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OCTOBER 1959

SAND-WEDGE POLYGONS (TESSELATIONS) IN THE McMURDO SOUND REGION, ANTARCTICA—A PROGRESS REPORT*

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ABSTRACT. A microrelief pattern of polygons 20 to 40 feet in diameter occurs in the perennially frozen ground of the glacier-free lowlands of the McMurdo Sound Region, Antarctica. They are similar to ice-wedge polygons in the Arctic except that they are in ground devoid of vegetation and are outlined by textural changes in the soil between the inter-polygonal furrow and the enclosed polygonal areas. The term "sand wedge" polygon is here proposed for this widespread phenomenon.

The sand wedge polygons occur in all surficial material, even in stagnant glacial ice. A transverse section across the interpolygon trench or furrow reveals a well-formed wedge-shaped mass of structureless fine to medium-grained sand extending 1 to 10 feet below the surface and ¼ to 4 feet wide at the top tapering down to a feather edge at the base. The sediments adjacent to the wedge have been deformed upward and may be vertical or overturned near the top of the wedge. The upwarping decreases downward.

Suggested origin of the polygons and sand wedges is similar to the origin of foliated ice wedges and polygons in the Arctic. Periodic contraction cracks in the perennially frozen ground around McMurdo Sound, cracks produced by the great change in temperature from summer to winter, are gradually filled with clean sand which filters down from above in the spring and summer. Repeated cracking and filling with sand produces a wedge-shaped filling.

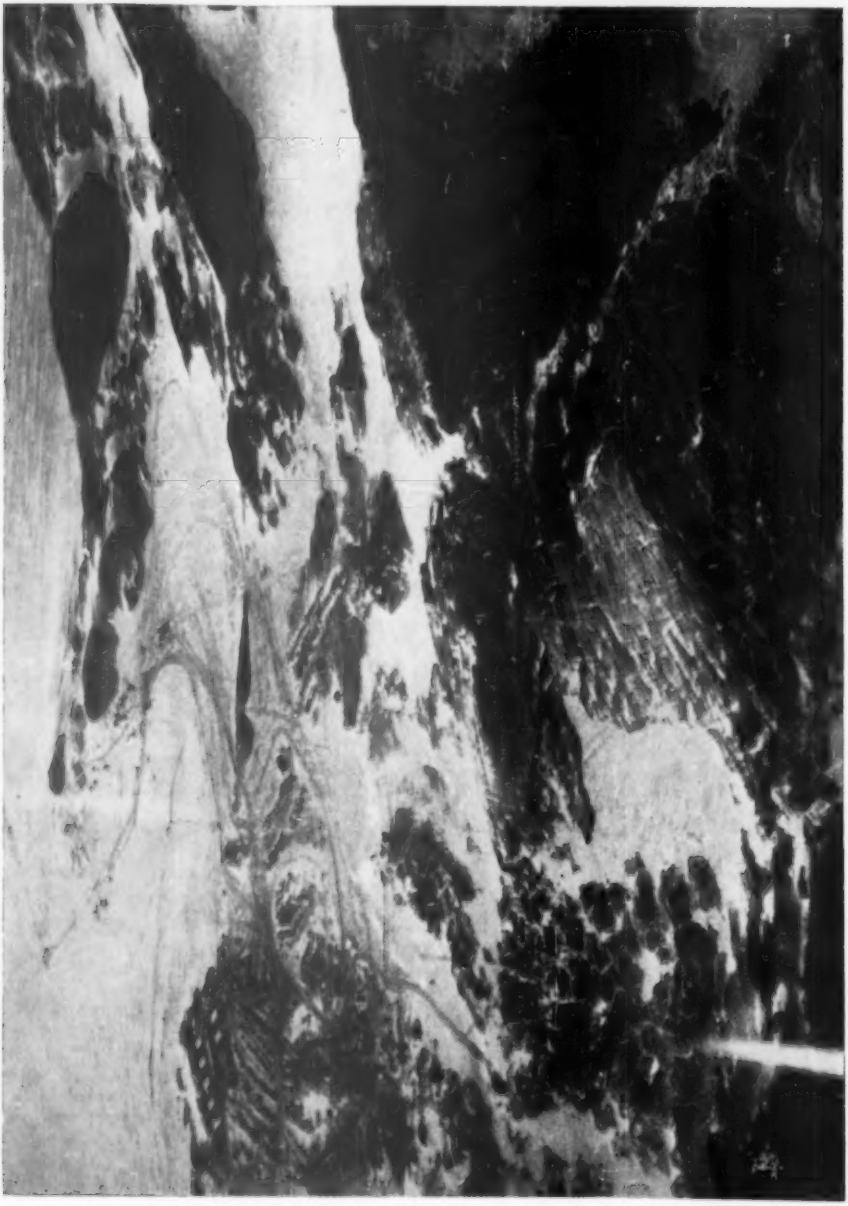
Perhaps not all sand wedges reported in temperate latitudes are replacements of ice wedges as generally believed, but may be original structures formed under an arid climate similar to that which exists at McMurdo Sound today.

A ubiquitous physical phenomenon in the ice-free lowlands of the McMurdo Sound Region, Antarctica, is a microrelief pattern of polygons or tesselations (Taylor, 1922) in the perennially frozen ground (fig. 1). The polygons are 20 to 40 feet in diameter and are formed by intersecting shallow furrows or trenches 3 to 18 inches deep. When a small amount of snow accumulates in the furrows the surface of the ground has the appearance of a large mosaic pavement. Such was succinctly stated by Taylor (1916, p. 115):

> The lowlands of loose black rock appear to be rolled by a steam roller, while the surface is broken by gutters from four to eight inches deep. These gutters marked out hexagonal and polygonal areas some twenty or thirty feet across. When a light snowfall had collected in the gutters, the valleys seemed to have been paved with black tiles united by white mortar.

A similar microrelief pattern (polygonal ground, tundra polygons, ice-wedge polygons) is widespread in northern North America, Russia, and other Arctic and subarctic areas but the polygons are outlined, in addition to the relief, by slight differences in the vegetation and by small linear ponds in the polygonal trenches. (Troll, 1944; Frost, 1950; Wiggins, 1951; Black, 1952b;

* Publication authorized by the Director, U. S. Geological Survey.



Péwé, 1954; Shumskiy, Shvetzov, and Dustovalov, 1955; Washburn, 1956; Britton, 1958a, 1958b). The pattern in Antarctica is in ground devoid of vegetation and, in the absence of snow-filled furrows, is outlined by textural changes in the soil between the furrows and the enclosed polygonal areas. Taylor (1922) mentioned the polygonal pattern (tesselations) but did not do detailed work or demonstrate a method of origin. The term "sand-wedge polygon" is here proposed for this phenomenon so widespread in the McMurdo Sound Region.

An opportunity to study the polygonal ground in Antarctica was provided by the United States National Committee of the International Geophysical Year and observations were made by the writer, assisted by Norman R. Rivard, in the ice-free areas east, south, and west of McMurdo Sound (fig. 2).

Fig. 2. McMurdo Sound Region, Antarctica.

from December, 1957 until February, 1958. The area around the Sound is composed of volcanic, intrusive, metamorphic, and sedimentary rocks. The lowlands are in large part covered with glacial deposits (Péwé, 1958b) and both alpine and outlet glaciers occur in this region (Péwé and Rivard, 1958).

The climate of the McMurdo Sound Region is rigorous and characterized by low temperatures and aridity. Scattered data from Scott's expeditions and from incomplete IGY observations indicate that the mean annual air temperature near sea level on the east side of McMurdo Sound is approximately 1°F. Data on the mean annual precipitation in the region is difficult to obtain be-

Fig. 1. Aerial view of polygonal ground (tesselations) in basaltic rubble, Hut Point, McMurdo Sound, Antarctica. Williams Naval Air Facility at upper left. U. S. Navy photo, 1956.

cause most of the precipitation is in the form of blowing snow and because the record is very short; however, on the east side of the Sound, the mean annual precipitation in inches of water is probably between 2 and 6. It is the opinion of the writer that the west side of the Sound is drier than the east but no quantitative data are yet available.

The sand-wedge polygons occur in all surficial material such as till, outwash, stream sand and gravel, lake sediments, deltaic sediments, rubble derived from bedrock by frost action, ablation moraine overlying stagnant glacial ice, and even in the stagnant glacial ice. The various types of sediments are firmly cemented by ice and are not dry permafrost. This microrelief pattern occurs on flat lands and on slopes up to 30 to 35 degrees inclination. Most polygons have raised edges 3 to 12 inches high and in many places the furrows are boulder lined. The very young polygons, which are on recent flood plains and the youngest stream terraces, do not have raised edges and the furrows are only 3 or 4 inches deep.

The absence of vegetation permits excellent observation of the structure and configuration of soil in the upturned edge of the polygon, a feature not often seen in the ice-wedge polygons of the Arctic and subarctic regions. A plan view reveals that the ground, which is covered with a deflation lag gravel, is turned up at the edge of the polygon and the linear and plate-like rocks stick out of the ground at steep angles on the ridge adjacent to the furrow (fig. 3). The width of the furrow or distance between raised edges of adjacent polygons ranges from 4 inches to 6 feet. The narrow bottom of the furrow is covered with either clean well-sorted wind-blown sand or cobbles and rock fragments that have rolled down into the furrow from the raised edges. In many places sand can be seen trickling down into a vertical crack that extends into the earth from the bottom of the furrow.

A transverse section (fig. 3) across the trench or furrow reveals a well-formed wedge-shaped mass of structureless fine- to medium-grained sand extending 1 to 10 feet below the surface and ¼ to 4 feet wide at the top tapering down to a feather edge at the base (fig. 3). The sand is cemented by ice and in many places a ¼ inch wide open crack extends downward 1 to 3 feet into the wedge from the upper surface of the permafrost. The sand wedge lies directly under the trench. The sediments adjacent to the wedge have been deformed upward and may be vertical or overturned near the top of the wedge. The upwarping or tilting of the sediments decreases downward and is absent near the base of the wedge. Cobbles or boulders in the parent material adjacent to the wedge crop out in the upturned edges of the sediments and most of the linear or plate-like rocks are oriented with their short axis perpendicular to the dip (fig. 3).

The writer has studied polygonal ground in Alaska for several years and has observed that in the Arctic and subarctic this microrelief pattern is, with rare exceptions (Hopkins and Karlstrom, 1955, pls. 39, 43), directly associated with underlying masses of foliated ground-ice, commonly termed "ice wedges" because of their wedge shape when exposed in transverse section (Leffingwell, 1915, 1919, figs. 16-25; Taber, 1943, pl. 17; Péwé, 1958a, fig. 2). Upon examination of photographs of tesselations in the Antarctic the

writer was prepared to find ice wedges and polygons similar to those in the Arctic; however, although the polygonal pattern and microrelief were similar, examination of many sections of the polygons in the McMurdo Sound Region revealed sand wedges in the same position occupied by ice wedges in the North. The sand wedges and the deformed strata adjacent to the wedge duplicate in great detail the features and field relations associated with the foliated ground-ice wedges as described from the North (Leffingwell, 1915, 1919; Black, 1951, 1952a, 1952b, 1954; Popov, 1955; Shumskiy, Shvetzov, and Dostovalov, 1955; Péwé, 1958a). The writer has noted large-scale contraction polygons without ice wedges in outwash gravel near Big Delta, Alaska, and Hopkins and Karlstrom (1955, pls. 39, 43) illustrate such polygons from the Bristol Bay area and Seward Peninsula, Alaska. Such "frost-crack" polygons (Hopkins and Karlstrom, 1955, p. 139) may contain a thin vein of silt or sand generally an inch or less in thickness. The bedding bends downward and extends across the "wedge" instead of bending upward adjacent to the wedges as in the sand wedges of Antarctica.

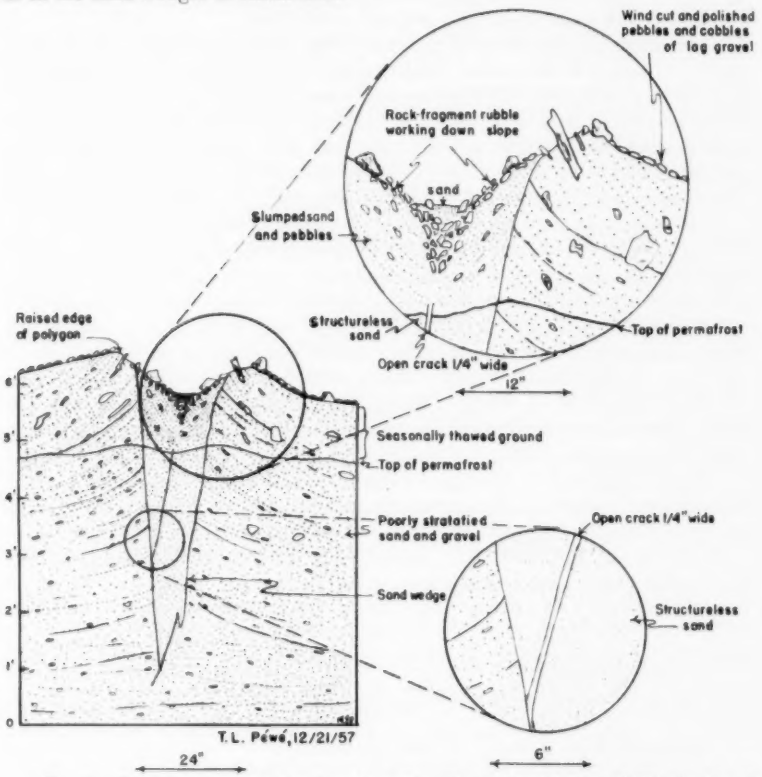


Fig. 3. Diagrammatic sketch of sand wedge, Taylor Dry Valley, McMurdo Sound, Antarctica.

The hypothesis entertained at this time for possible origin of the polygons and sand wedges is similar to the hypothesis for origin of foliated ice wedges and polygons in the North proposed by Leffingwell (1915, 1919) and supported by Black (1952a) and Péwé (1952). Leffingwell postulates (1919, p. 205-212) that foliated ground-ice forms in periodic contraction cracks in the perennially frozen ground—cracks which are produced by the great changes in temperature from summer to winter. The intense stresses generated by a drop in ground temperature result in the formation of a polygonal network of contraction cracks. Hoar-frost, ice crystals, and snow melt water periodically fill or partly fill the cracks to produce a narrow vein of ground-ice in the permafrost. This ice vein is not destroyed by the expansion of the ground in summer and provides a zone of weakness for repeated cracking and ice growth in subsequent winters. Eventually a polygonal network of massive foliated ice masses (wedges) exists in the ground. In the arid ice-free region around McMurdo Sound such contraction cracks are gradually filled with clean sand which filters down from above in the spring and summer. Repeated cracking and filling with sand produces a wedge-shaped filling.

An alternate hypothesis is that ice wedges formerly existed but later disappeared, and that the present sand wedges are "fossil" ice wedges or pseudomorphs after ice wedges. This suggestion can be ruled out because: (1) On many occasions sand was observed trickling down the thermal contraction ground crack in old large wedges and small new wedges. (2) Sand wedges were observed in stagnant glacial ice, a medium which would disappear along with the ice wedges if the later were to melt. (3) The region is so extremely arid, especially on the west side of the Sound, that there is little snow at low levels and most of this disappears by evaporation or sublimation into the atmosphere. (4) No evidence is available to support the suggestion that permafrost and ice wedges disappeared or thawed to produce "fossil" ice wedges.

One conclusion that can be drawn at this early stage of this study is that sand or sediment wedges reported from temperate latitudes in North America (Horberg, 1949; Schafer, 1949; Black, 1957; Wilson, 1958) and Europe (Lotze, 1932; Cailleux, 1948; Dylik, 1951; Dimbleby, 1952; Cailleux and Taylor, 1954; Jahn, 1955; Dutkiewicz, 1956; Dylik, 1956a, 1956b; Dylikowa, 1956; Golab, 1956; Klatka, 1956; Pierzchalko, 1956; Sekyra, 1956) possibly should be reevaluated in the light of discoveries in Antarctica. Perhaps not all the sediment wedges in temperate latitudes, especially if composed of sand, are replacements of ice wedges which formed in a climate such as now exists in the Arctic of Asia and North America but possibly may be original structures formed under an arid climate similar to that which exists at McMurdo Sound today.

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ANORTHITE CONTENT OF PLAGIOCLASE IN BASIC AND PELITIC CRYSTALLINE SCHISTS AS RELATED TO METAMORPHIC ZONING IN THE USU MASSIF, TIMOR

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ABSTRACT. Systematic sampling of basic and pelitic crystalline schists in the Usu massif has yielded information relating to the anorthite content of plagioclase in the greenschist and in the lower-grade portion of the almandine amphibolite facies. Frequency curves in which frequency of occurrence is plotted against anorthite content have been constructed for basic and for pelitic schists.

1. The frequency of occurrence of plagioclase between An_5 and An_{20} in basic rocks was found to be remarkably low, in contrast with the high frequencies for plagioclases on both sides of this interval. In the field the values of the interval An_5 to An_{20} occupy a relatively narrow zone. During metamorphism the transition from An_5 to An_{20} apparently took place in a narrow pressure-temperature field.

2. The general shape of the frequency curve for pelitic rocks has the same characteristics as that for basic rocks. Differences appear in values between An_{10} and An_{30-35}, which are generally higher, and in values above An_{30-35}, which are commonly lower than those in basic rocks.

3. The An_{10} boundary between the greenschist facies and the almandine amphibolite facies as advocated in the facies classification of Fyfe, Turner and Verhoogen (1958) proves to be a sharp one in the Usu massif. The abrupt transition around An_{10} characterizes the facies boundary which can be traced in the field in great detail. For the determination of the An_{10} isograd both basic and pelitic rocks can be used, as the two frequency curves intersect in this point.

4. The An_{20} isograd used by some writers (Ramberg, 1949, 1952; Barth, 1952) as a facies boundary appears as an arbitrary point on a smooth part of the curve, and is, therefore, far less distinctive than the An_{10} isograd.

INTRODUCTION

The Usu massif is located NE of the town of Kefamenanu in the central part of western (Indonesian) Timor. The massif which measures about 23 square miles is a tectonic window, surrounded by sedimentary rocks of the Sonnebait tectonic unit (de Waard, 1957a). The area consists of alternating basic and pelitic schists of low to medium-grade of regional metamorphism. The age of the orogeny and metamorphism is pre-Alpine, presumably Variscan.

Earlier field investigations in the Lalan Asu massif, 60 miles southwest of the Usu area, revealed changes in the An content of the plagioclase over relatively short distances reflecting the succession of narrow zones of metamorphism. The intricate pattern of isograds has been attributed (de Waard, 1957b) to intense late-metamorphic deformation of a more regularly distributed prekinematic metamorphic zoning.

In order to investigate more closely the plagioclase composition in relation to metamorphic zoning the Usu massif was mapped and sampled as systematically as possible. The resulting network of 204 stations, about 500 to 750 meters apart, yields an average sample of the range of metamorphic conditions in the area, and as well reveals the behavior of the An content of plagioclase under those conditions, and enables the construction of metamorphic isograds and plagioclase isopleths in some detail. Basic and pelitic crystalline schists alternate in sheets of greatly varying thickness. Basic schists were found in

168 out of the 204 stations, and pelitic schists occur in 107 stations. In 71 stations it was possible to collect samples of both the basic and the pelitic schist which enables a close comparison of these rock types as formed under identical metamorphic conditions.

The purpose of this paper is to illustrate the behavior of the An content of plagioclase in basic schists under low- and medium-grade metamorphic conditions of the Usu massif, and to draw attention to systematic differences in the plagioclase composition between pelitic and basic schists formed under the same metamorphic conditions. The plagioclase determinations were carried out with the universal stage using the method and curves of Turner (1947).

ROCK TYPES, METAMORPHIC FACIES AND ROCK TEXTURES

A brief description of the mineral assemblages of the rocks, the grades of metamorphism and rock textures will be given to illustrate the range of physical conditions and of chemical composition represented in the area.

A. The basic crystalline schists consist of greenschists and amphibolites; with increasing grade of metamorphism the following mineral associations are found; minor constituents are noted in parentheses:

albite-chlorite-actinolite-epidote (quartz, calcite, muscovite, stilpnomelane, sphene);

biotite-albite-chlorite-actinolite-epidote (quartz, calcite, muscovite, stilpnomelane, sphene);

epidote-albite-hornblende (quartz, calcite, sphene);

oligoclase-hornblende (epidote, quartz, calcite, sphene, biotite, garnet);

andesine-hornblende (epidote, quartz, sphene, biotite, garnet, diopside);

labradorite or bytownite-hornblende (epidote, quartz, garnet, biotite, sphene).

B. The pelitic crystalline schists include phyllites, mica schists and gneisses:

muscovite-chlorite-albite-quartz (epidote, calcite, tourmaline);

biotite-muscovite-chlorite-albite-quartz (epidote, calcite, tourmaline);

garnet-biotite-muscovite-albite-quartz (chlorite, epidote, tourmaline);

garnet-biotite-muscovite-oligoclase-quartz (chlorite, epidote, hornblende, tourmaline);

garnet-biotite-muscovite-andesine-quartz (chlorite, epidote, hornblende, kyanite, tourmaline).

Following the revised facies classification of Fyfe, Turner and Verhoogen (1958) the assemblages of the Usu massif indicate a regional metamorphism from the lowest grade of the greenschist facies to medium almandine amphibolite facies. The transition at An_{10} from albite to oligoclase in the assemblages represents the high temperature boundary of the greenschist facies. The uppermost grade of metamorphism in the area stays well within the almandine amphibolite facies. Although scarcity of characteristic minerals hampers the recognition of subfacies everywhere, there are indications that metamorphism reached the middle one of the three subfacies now recognized, viz., small grains of kyanite in part of the almandine mica schists, large diopside crystals in some of the epidote-andesine amphibolites, and basic andesine, labradorite and bytownite in a number of amphibolites.

Among the metamorphic textures are found both purely crystalloblastic and phyllonitic textures with transitional textures in between. The range of textures, from rotated, helicitic porphyroblastic to strongly oriented nematoblastic and lepidoblastic, and to phyllonitic textures with or without retrogressive effects, indicates an akinematic, progressive regional metamorphism which graded into synkinematic metamorphism, and which concluded with a locally retrogressive phyllonitization. Although phyllonitization of some degree is fairly common in the area, only some rocks show obvious mineralogical evidence of disequilibrium. Thus plagioclase may appear in relics having a composition different from that of grains in an enclosing granulated mass. It may occur also in zoned crystals in which the An content changes gradually or irregularly. The change from core to rim is commonly a decrease in the An content, though the opposite also frequently occurs (from core to rim to outer edge, e.g., 37-41-51, 34-52-70; or mixed zones, e.g., 34-37-30).

PLAGIOCLASE IN THE BASIC ROCK SERIES

The 168 sample localities in which plagioclase-bearing greenschists and amphibolites are found are fairly equally distributed throughout the area.

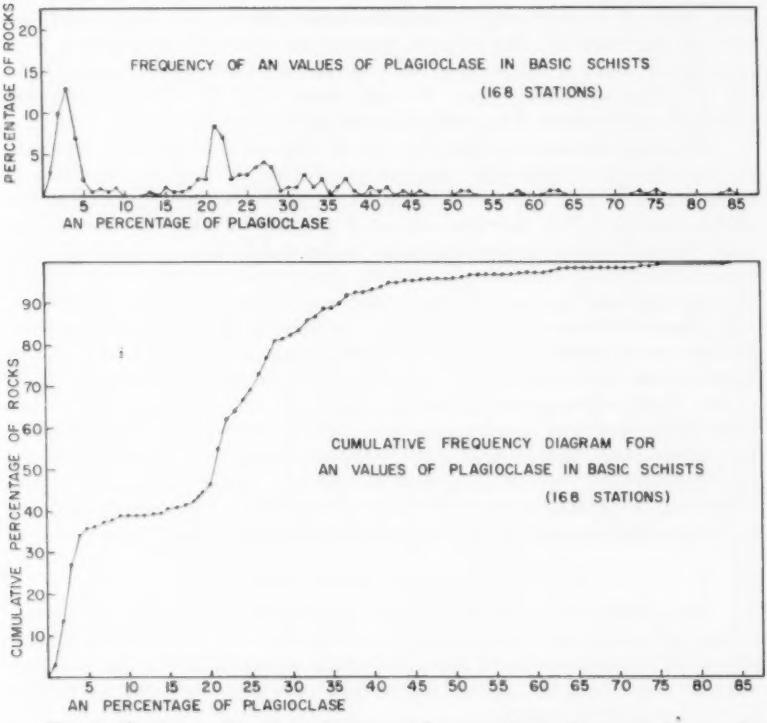
PERCENTAGE OF ROCKS
FREQUENCY OF AN VALUES OF PLAGIOCLASE IN BASIC SCHISTS
(168 STATIONS)
AN PERCENTAGE OF PLAGIOCLASE
CUMULATIVE PERCENTAGE OF ROCKS
CUMULATIVE FREQUENCY DIAGRAM FOR
AN VALUES OF PLAGIOCLASE IN BASIC SCHISTS
(168 STATIONS)
AN PERCENTAGE OF PLAGIOCLASE

Fig. 1. Frequency diagram and corresponding cumulative frequency diagram for basic schists arranged according to anorthite content of plagioclase.

Curves based on the frequency of the An values, therefore, reflect in a general way the behavior of the plagioclase composition under the various metamorphic conditions represented in the massif.

Figure 1 illustrates the frequency of the various An values of the plagioclase in basic schists. Albite up to An_5 appears frequently (36%) in contrast with albite-oligoclase between An_5 and An_{20} (9%). A second peak is reached in the oligoclase range of An_{20-25}, beyond which the frequency gradually decreases to about An_{45} with occasional occurrences of plagioclase up to An_{84}. The break in the frequency between An_5 and An_{20} is of particular interest. Although virtually all plagioclase members between An_0 and An_{84} are present, those between An_5 and An_{20} are relatively rare as compared to the high frequency of An values on both sides of this gap. The cumulative frequency diagram illustrates this particularly well. The horizontal part of the curve which represents the An_{5-20} interval forms an obvious interruption in an otherwise fairly close approximation of a probability curve.

The shape of the frequency curves reflects the occurrence of a narrow zone of transitional An values between relatively large areas of albite-bearing rocks and rocks with plagioclase above An_{20}. The distribution of the An values on the map, figure 2, is elucidated by isopleths drawn between the sample localities with 10 An intervals. The isopleth pattern resembles that of contours on a topographic map. A flat lowland of albite-bearing rocks extends toward the foot of a steep cliff of the An_{5-20} interval, above which there is an undulating plateau of values above An_{20} with occasional peaks up to An_{84}.

The field distribution of the An values of plagioclase largely coincides with and parallels the metamorphic isograds as constructed from the occurrence of index minerals and mineral assemblages in basic and pelitic rocks of the sample localities. The isograds suggest a gradual increase in metamorphic grade from the lower greenschist facies to the middle almandine amphibolite facies. Under those conditions a similarly gradual increase of the An content of plagioclase may be expected in accordance with the equilibrium reaction for plagioclase in the presence of epidote as described by Ramberg (1944, 1949). Epidote as the supplier of lime in the plagioclase reaction is nearly always present in the basic rocks of the Usu massif (97% of the greenschists and 82% of the amphibolites contain epidote). There are no indications that abrupt chemical or structural changes may have affected the regional distribution of the various plagioclase members. For a reason other than structural or chemical inhomogeneity the An_{5-20} interval as based on field evidence in this particular area appears to differ from Ramberg's theoretical curve which predicted a more gradual increase of the An count with increasing metamorphic grade.

PLAGIOCLASE IN THE PELITIC ROCK SERIES

In 71 of the 107 localities in which plagioclase-bearing pelitic schists are found, basic schist samples also have been collected. This enables a close comparison of the An contents of basic and pelitic schists formed under the same metamorphic conditions, as the samples have been taken as a rule not farther apart than 10 or 20 meters.

In figure 3 the An values of plagioclase in pelitic schists are plotted

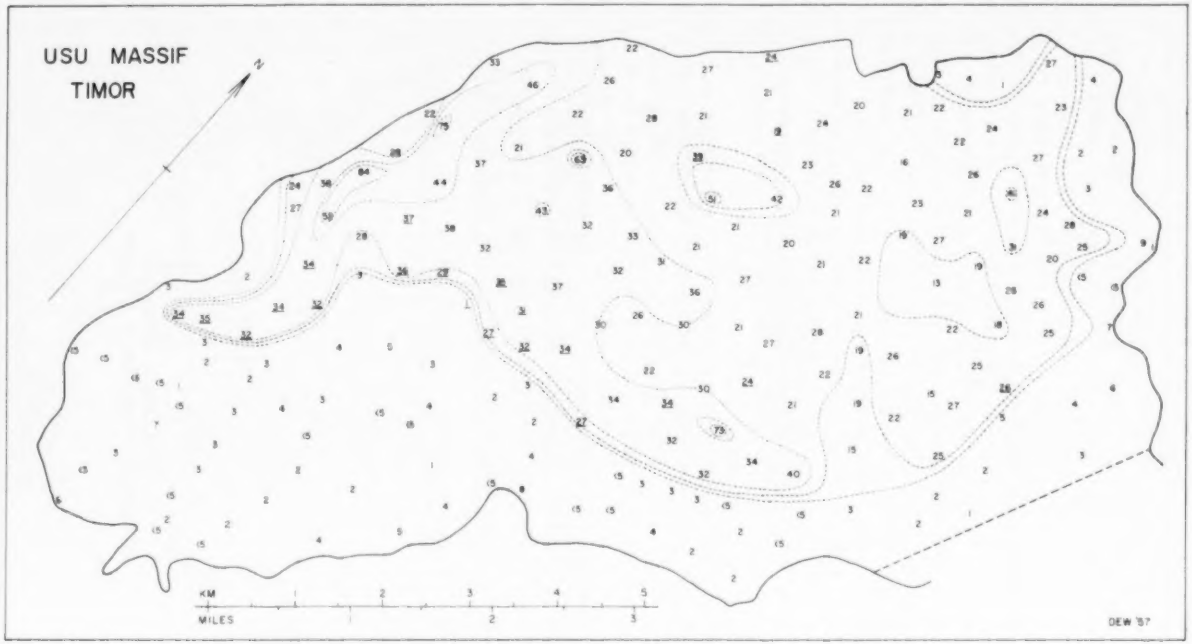

Fig. 2. Anorthite percentages and anorthite isopleths in the Usu massif. Underlined numbers indicate the occurrence of pelitic schists only. Isopleths are drawn every 10 An up to An_{50}.

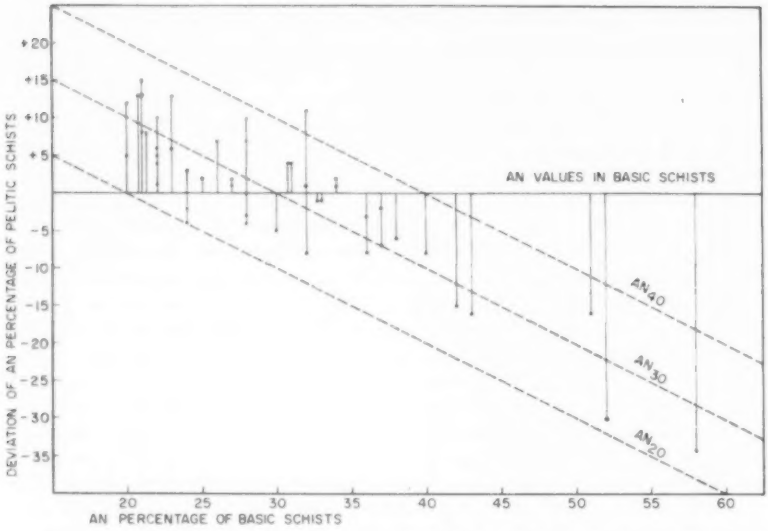

Fig. 3. Comparison between An content of plagioclase in basic (horizontal line) and that in associated pelitic schists (dots). The distance between the dot and the horizontal line represents the difference in An content between the pelitic schist and the associated basic schist.

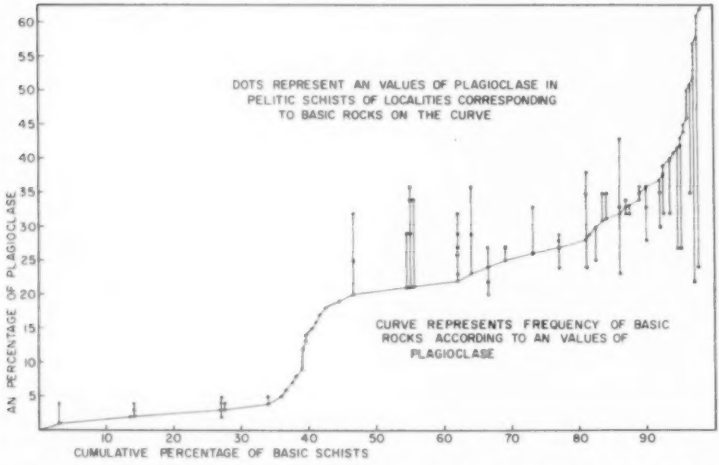

Fig. 4. Cumulative frequency curve for An content of plagioclase in basic schists. Dots represents An content of plagioclase in pelitic rocks, tied vertically to the An content of plagioclase in associated basic schists on the curve.

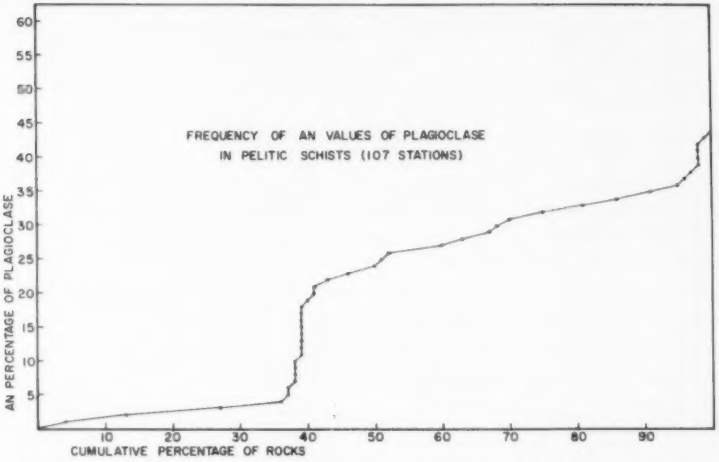

Fig. 5. Cumulative frequency curve for An content of plagioclase in pelitic schists.

against those for associated basic schists. The diagram shows that the An values of the two rock types may differ as much as 34 percent from each other. The differences, however, are not at random. There is a distinct tendency for the An values to be higher in pelitic than in the basic schists for An values between 20 and 30 to 35, and lower than in the basic schists for An values above 30 to 35. In pelitic rocks the An value if in excess of 20 tends to lie between 25 and 35, i.e., between the dotted lines of figure 3. Near An_{30} to An_{35} the plagioclase of the pelitic and basic metamorphic rocks in this area appears to have approximately the same An content.

In figure 4 the deviation of An values of pelitic schists is plotted against the An frequency curve for associated basic schists. Albite-bearing pelitic schists show negligible differences in composition as compared with the basic schists of the same locality. Above An_{20} in basic schists pelitic rocks tend to deviate toward a zone situated largely between An_{25} and An_{35}.

This deviating trend is also clearly expressed in the cumulative frequency diagram of An values in pelitic schists, figure 5, which in its general aspects has the same characteristics as the curve for the basic schists. In detail the curves for pelitic and for basic schists are not mutually comparable, however, since the data on which the curves are based are partly derived from different localities. As the pelitic sample localities are fairly equally distributed over the area the general outline of the curve may be considered to reflect the average behavior of the An content in plagioclase of the pelitic schists. The abrupt transition from An_5 to An_{20} and the gradual increase in An content above An_{20} are common to both frequency curves. The curves differ as regards the values between An_{20} and An_{30} (higher An in pelitic than in basic schists), and above An_{35} (generally lower An in pelitic than in basic schists). The highest value found in pelitic schists is An_{44}, as compared to An_{84} in basic schists.

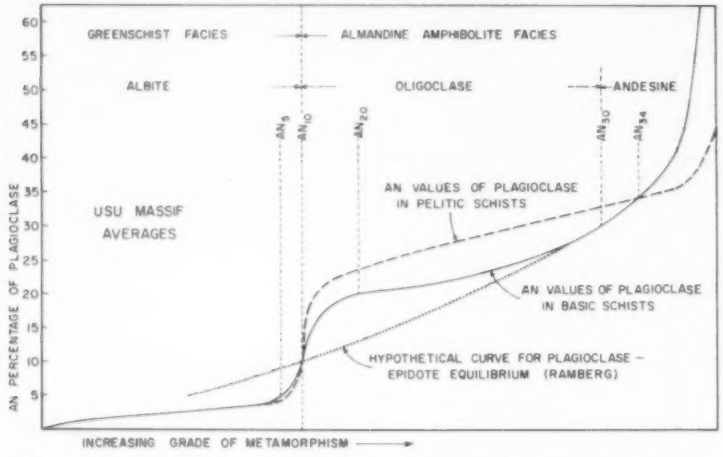

Fig. 6. Diagrammatic curves representing the relation of An content of plagioclase to grade of metamorphism. The scale of metamorphic grade is hypothetical and not necessarily uniform.

SUMMARY AND CONCLUSIONS

Systematic sampling of schists in the Usu massif yielded a detailed pattern of isopleths based on An values of plagioclase. The An_{10} isopleth marks the isograd separating the greenschist facies from the almandine amphibolite facies. As there is no obvious chemical or structural heterogeneity in the area which may have seriously affected the An content of the plagioclase, the grade of metamorphism must have been primarily responsible for the plagioclase equilibrium which is assumed to have been attained in the manner as represented in Ramberg's (1949) theoretical diagram showing the change of composition of plagioclase in equilibrium with epidote. The frequency curves of the An values of plagioclase as based on an equal distribution of the samples throughout the massif provide a means to check the general trend of the plagioclase reaction in this particular area against the theoretical curve (fig. 6).

1. In basic schists the transition from albite to calcic plagioclase with increasing grade of metamorphism is not so gradual as the theoretical curve suggests. The break in the frequency curve between An_5 and An_{20} reflects the relatively rare occurrence of plagioclase of this kind. In the field the isopleths of An_{10} and An_{20} are generally found very close to each other or are superimposed. The transition from about An_5 to about An_{20} is an abrupt one. This particular behavior of the An content as described here for the Usu area should probably not be regarded as exceptional. Fyfe, Turner and Verhoogen (1958, p. 218) write: "With rising grade of metamorphism there is a sudden jump from albite (An_{0-7}) to oligoclase-andesine (An_{15-30}), in quartzo-feldspathic, semipelitic, and basic schists."

2. In pelitic schists the frequency curve of the An values is in its general trend similar to that of the basic schists. Here also there is a general lack of

rocks with plagioclase between An_5 and An_{20}, suggesting an abrupt transition from albite to more calcic plagioclase in the same interval as in the basic schists. The noteworthy difference is the relatively higher An values between An_{20} and An_{30-35} and lower values above An_{30-35} as compared to the plagioclase in basic schists. This is probably the result of initial differences in the chemical composition between the two rock types. Above An_{30-35}, the lower values for An in plagioclase of pelitic schists probably reflect a lower content of CaO in pelitic as contrasted with basic schists. Between An_{20} and An_{30-35} the higher values for An in plagioclase of pelitic schists are attributed to the influence of coupled reactions (Fyfe, Turner and Verhoogen, 1958, p. 24, 151-153). On a pressure (P_{H_2O})-temperature diagram the exact position of the curve of univariant equilibrium for the reaction

$$\text{albite} + \text{clinozoisite} + \text{alumina} \rightleftarrows \text{plagioclase}^1 + \text{water}$$

will depend upon free energy changes relating to more or less simultaneous reactions in which other minerals are formed. Thus the temperature and pressure at which oligoclase (An_{10-15}) first appears will be influenced by the formation of garnet from chlorite in pelitic schists and by chemical changes in the actinolite-hornblende series in basic schists. While the sudden increase of anorthite from about An_5 to An_{15} in both types of rocks marks a sharply defined isograd, the exact temperature and pressure at which some higher value (e.g., An_{20}) is reached is not necessarily the same in the pelitic as in the amphibolite paragenesis.

3. The An_{10} isopleth in the field represents, at least in the Usu area, a very distinct facies boundary between the greenschist facies and the almandine amphibolite facies. The interval between An_5 and An_{20} occurs so abruptly and occupies such a narrow zone in the field, that the boundary between the greenschist and almandine amphibolite facies can be traced in great detail. As the An_{5-20} interval occurs identically in the plagioclase of pelitic and basic schists both rock types can be used indiscriminately for the determination of the facies boundary.

4. The An_{30} isopleth has been advocated as the facies boundary between the epidote amphibolite and amphibolite facies by Ramberg (1949, 1952), Barth (1952) and others as one way out of the ambiguous facies boundary in the original facies classification of Eskola (1914, 1939). The An_{10} boundary is favored by Turner (1948), and Turner and Verhoogen (1951), and is recently emphasized in the revised facies classification of Fyfe, Turner and Verhoogen (1958) in which the epidote amphibolite facies is considered the upper subfacies of the greenschist facies. Although the An_{30} boundary was previously also used by the present writer (1957b) for the Lalan Asu area, the present data from the Usu area unquestionably indicate that the An_{10} boundary has quite a different character and is by far the more natural one. The An_{30} isopleth appears merely as an arbitrary point on a smooth part of the curve, in contrast with the sudden dip of the curve around An_{10}. This strongly justifies using the An_{10} isopleth as the well-marked upper boundary of the greenschist facies.

[1] To define the unique shape of the curve some specific composition of plagioclase must be arbitrarily selected, e.g., An_{20}.

When An_{30} isopleths are used for subdivision of the almandine amphibolite facies it should be realized that identical An values in pelitic and in basic schists have almost the same significance. In this area the average curve for pelitic rocks intersects the curve for the basic rocks at An_{34}. There are, however, considerable variations in An values in this part of the curve, and these, having here a relatively greater effect than those near the An_{10} isograd, will tend to increase the vagueness of the An_{30} boundary as an isograd.

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PHASE RELATIONS IN THE SYSTEM Al_2O_3–H_2O AT HIGH TEMPERATURES AND PRESSURES*

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ABSTRACT. The system Al_2O_3—H_2O has been studied in the temperature range 100-600°C and at H_2O pressures ranging from 5,000 to 40,000 atmospheres. At 40,000 atmospheres pressure the maximum temperature at which the trihydrate, gibbsite, is stable is 295°. The monohydrate, diaspore, is stable at 40,000 atmospheres of H_2O to a temperature of approximately 590°. The relations of the polymorphs, diaspore and boehmite, are discussed and it is concluded that all boehmite forms metastably. The approximate slope of the boehmite-diaspore boundary, computed from thermodynamic data, indicates that diaspore in bauxite and clay deposits may have formed at atmospheric pressures and temperatures, a familiar conclusion from field observations.

INTRODUCTION

At least four minerals in the system Al_2O_3—H_2O are known. Knowledge of the phase relations in this system may provide evidence of temperatures and water pressures in a variety of geological environments. For example, the interpretation of the origin of the minerals of lateritic deposits, the chief sources of aluminum ore, depends upon an understanding of the phase relations in this system.

PREVIOUS WORK

The system Al_2O_3—H_2O has been widely discussed. Early investigations have been summarized by Fricke and Hüttig (1944). Most of the early work consisted of differential thermal analysis studies and served only to give the maximum temperatures at which mineral phases may exist metastably at 1 atmosphere.

Laubengayer and Weisz (1943) attempted to determine equilibrium positions of phases in this system under hydrothermal conditions. In a series of hydrothermal experiments they determined the sequence of phases in the system Al_2O_3—H_2O in contact with water under its own vapor pressure, up to the critical temperature of water. They also made a few experiments slightly above the critical temperature. In their runs, lasting 50 to 150 hours, gibbsite was converted to the monohydrate, boehmite, at approximately 150°C; bayerite was converted to boehmite at essentially the same temperature; disapore appeared from boehmite at approximately 275°, and corundum appeared from diaspore at pressures of 400 bars and at temperatures slightly over 400°C. Laubengayer and Weisz suggest as a possibility that boehmite is never a stable phase in the Al_2O_3—H_2O system but that the rate of change of boehmite to diaspore is exceedingly slow. These authors point out they were only certain of equilibrium in the transition of diaspore to corundum plus water. This particular equilibrium was established from either side of the transition point.

The later study of Ervin and Osborn (1951) closely confirms the results of Laubengayer and Weisz and extends these studies to higher pressures. Ervin and Osborn, under somewhat similar hydrothermal conditions, determined that gibbsite is converted to diaspore at approximately 300°C at vapor pres-

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sures equal to or greater than that of liquid water at this temperature, and that diaspore is dehydrated to corundum at approximately 400°C at pressures of 65-650 bars of water. In an extended series of comparable experiments the writer has produced essentially identical results and confirmed both of these earlier studies.

A good deal of evidence, internal to the phase diagrams, suggests that the results of Laubengayer and Weisz, the results of Ervin and Osborn and the writer's results, from hydrothermal studies in this temperature and pressure region, do not represent equilibrium determinations of the phase boundary positions. Only the boundary between diaspore and corundum has been reversed; that is, equilibrium has been approached from both directions in all three investigations. The gibbsite-boehmite transition and the boehmite-diaspore transition are both exceedingly sluggish. Further, in many low temperature runs nucleation of diaspore is necessary.

The slope and shape of the boehmite-diaspore boundary as reported by Ervin and Osborn is a most unusual shape for a solid-solid transition. The slope of a transition on a pressure-temperature plot, dp/dt, is equal to $\Delta s/\Delta v$. However, as compressibility and coefficient of thermal expansions of both phases is essentially identical, changes in volume with pressure and temperature changes on a phase, makes very little difference to the value of Δv; Δv is normally independent of pressure and temperature. Similarly, as both minerals are close to their Debye temperatures the value of Δs is relatively independent of the temperature of the reaction. Therefore, we might well expect the boundary separating the field of boehmite from diaspore to be essentially a straight line. Most experimentally studied solid-solid transitions have a straight line boundary over an appreciable range of temperature and pressure. Thompson (1955) has discussed these transitions and states, "It can be shown that the ratio $\Delta s/\Delta v$ for solid reactions may be regarded as constant." Yoder and Weir (1951) have demonstrated theoretically that for the specific reaction, nepheline plus albite equals 2 jade, providing solid solution does not intervene, the slope of this boundary will be constant. Furthermore, a fairly good estimate can be made as to the sign of the slope, dp/dt, for the boehmite-diaspore transition. Most solid-solid transitions have positive slopes, that is, the higher the temperature of the transition, the higher the pressure required. This is particularly true for all the transitions where there is a fairly large percentage change in the gram-atomic volume. The gram-atomic volume of diaspore, based on computations from x-ray determinations of the size of the unit cell, is 4.46 cc, that of boehmite is 4.83 cc. This is roughly an 8.5 percent change in the gram-atomic volume and should be accompanied by a rather large positive entropy change. Therefore we might expect that the boundary on a p-t plane separating diaspore from boehmite should be reasonably close to a straight line and have a positive slope.

The phase diagram of Ervin and Osborn, however, shows in general a boundary with negative slope separating the stability fields of boehmite and diaspore. Further, in the temperature region between 300 and 400°, the slope of the boundary is essentially zero, whereas at temperatures of around 270° the slope of the boundary is essentially infinite. This implies that the value

$\Delta s/\Delta v$ changes from essentially infinite to essentially zero in a temperature interval of less than 50°.

The inconsistencies between the predicted and the observed slopes of the boundaries separating the phases suggests strongly that a rate process rather than an equilibrium boundary is involved, as Laubengayer and Weisz suggested. Further, the geologic field evidence strongly suggests the diaspore deposits which have been thoroughly studied in this country, particularly the ones in Missouri and North Central Pennsylvania, were certainly formed at modest temperatures and pressures. Most of the students of these deposits are emphatic on this point. Thus both the field and laboratory evidence and the thermodynamic inconsistencies in the past work suggest that the stability fields of these minerals have not been adequately delineated.

EXPERIMENTAL DATA

Two types of apparatus were used in the present attempt to delineate the stability fields of boehmite, diaspore and corundum: conventional hydrothermal apparatus and piston-anvil type apparatus.

All runs at high pressures, above 5000, bars were made in a piston-anvil type of apparatus as described by Bridgman (1935) and as modified by Griggs and Kennedy (1956). In this apparatus the sample is a wafer of material inside a stainless steel washer between two carboloy pistons.

Vapor pressures of hydrated phases can readily be determined in such

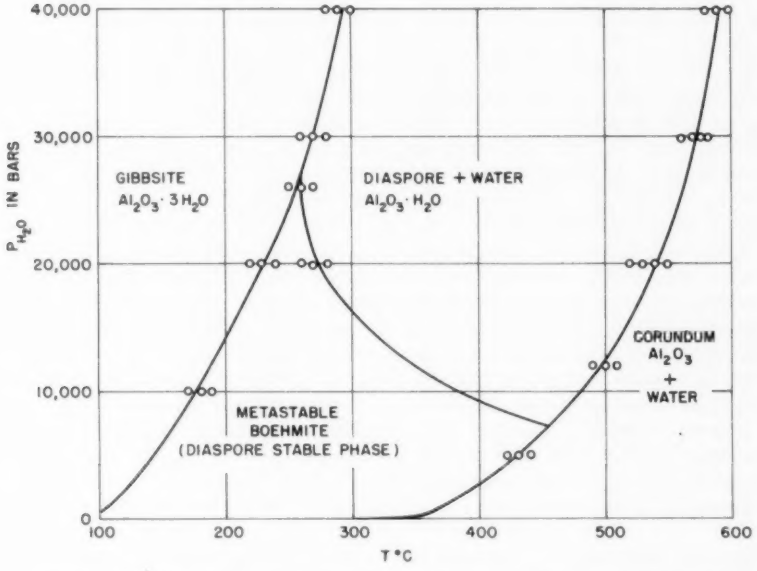

FIGURE 1. PHASE RELATIONS AT HIGH H_2O PRESSURES IN THE SYSTEM $Al_2O_3 \cdot H_2O$

Fig. 1. Phase relations at high H_2O pressures in the system Al_2O_3—H_2O.

an apparatus, for it the vapor pressure of the hydrated phase is less than piston pressure at a given temperature, the hydrated phase will persist during the run and can later be identified by x-ray or optical methods. If the vapor pressure of the hydrated phase at a given temperature is greater than applied piston load, the vapor will lift the pistons and escape, and anhydrous assemblages result. With this trial-and-error method the vapor pressure curves of various hydrates can be determined to very high temperatures and pressures. In this investigation, for instance, the vapor pressure of gibbsite was determined to 295°C where the H_2O pressure is approximately 40,000 bars (approximately 600,000 psi). The vapor pressure of diaspore reaches 40,000 bars at approximately 590°C.

Several hundred runs in the piston-anvil apparatus have been made in the system Al_2O_3—H_2O. Those used in delineating the phase boundaries are listed in table 1 and are plotted in figure 1.

The starting substance in all these runs was a mixture of roughly equal

TABLE 1

Runs with piston-anvil device in system Al_2O_3—H_2O

Temperature C°	Pressure in Kilobars	Products
420	5	Boe
430	5	Boe + Cor
440	5	Cor
170	10	Ba + Gi
180	10	Boe
190	10	Boe
220	20	Ba + Gi
230	20	Ba + Gi + Boe
240	20	Boe
260	20	Boe
270	20	Boe
280	20	Boe + Di
520	20	Di
530	20	Di
540	20	Di
550	20	Cor
250	26	Ba + Gi
260	26	Ba + Gi + Di + Boe
270	26	Boe + Di
260	30	Ba + Gi
270	30	Ba + Gi + Di + Boe
280	30	Boe + Di
560	30	Di
570	30	Di
575	30	Cor
580	30	Cor
280	40	Ba + Gi
290	40	Ba + Gi
300	40	Di
580	40	Di
590	40	Di
600	40	Cor

Gi = Gibbsite
Ba = Bayerite
Di = Diaspore
Boe = Boehmite
Cor = Corundum

parts of gibbsite and bayerite. The gibbsite was Baker's analyzed "aluminum hydroxide". The bayerite was prepared by neutralizing $AlNO_3$ with NH_4OH. The resulting gel was thoroughly washed and dried at 80°C. X-ray examination of the dried material showed it to be largely if not entirely crystalline bayerite.

All runs in the apparatus were for approximately one hour. Surprisingly, reaction in a sample compressed between pistons and brought to a selected temperature usually proceeds as far as it will go within the first few minutes at the chosen temperature and pressure. Runs of several days length have been tried, but normally in this apparatus reactions proceed within the first few minutes or not at all. Presumably the drastic change of reaction rate with time arises from the fact that a large amount of strain energy is stored in the sample on initial compression and is quickly relieved as the mineral recrystallizes. This initial strain energy appears to greatly accelerate reaction rate in the system, but the rate of conversion of one phase to another quickly declines as recrystallization takes place.

It is clear, from examination of the experiments listed in table 1, that in the vast majority of runs equilibrium was not achieved. Bayerite, precipitated metastably at one atmosphere from aluminum gel, persisted in all runs in the gibbsite field. This was most remarkable and unexpected as bayerite can be readily converted to gibbsite at one atmosphere pressure by heating at temperatures of approximately 200°C. Apparently at high pressures the viscosity of the bayerite is greatly increased and conversion to the more stable gibbsite inhibited. Bayerite has no known field of stability at one atmosphere pressure and is always synthesized metastably. However, it is somewhat denser than the stable trihydrate, gibbsite. The gram-atomic volumes of the two phases are respectively 4.47 cc and 4.57 cc, and at sufficiently high pressures we might expect bayerite to become a stable phase. However, conversion of gibbsite to bayerite was not observed. At temperatures above 260° and H_2O pressures of the order of 26 kb gibbsite dehydrates directly to mixtures of diaspore and boehmite. Below 260° boehmite is the phase formed on dehydration of the mixture of gibbsite and bayerite. It is clear that diaspore is the stable phase and the boehmite only forms metastably in most of the large area labeled "Metastable Boehmite" in figure 1. The transition of boehmite to diaspore is very sluggish. Long continued hydrothermal runs at low temperatures will convert boehmite into the diaspore in much of this temperature-pressure region of "Metastable Boehmite". The boundary separating the fields of metastable boehmite and diaspore in figure 1 is essentially the line of constant rate of conversion in the particular apparatus used. The rate of transition of boehmite to diaspore is speeded up both by higher pressures and higher temperatures. The effect of pressure on the rate of conversion of metastable bayerite to gibbsite and of metastable boehmite to diaspore is of opposite sign. High pressure inhibits the bayerite-gibbsite reaction but speeds up the boehmite-diaspore reaction.

The diaspore to corundum plus water boundary, shown in figure 1, is a sharp transition, readily delineated, and presents no problems.

In addition to the high pressure runs made in the piston-anvil device, a

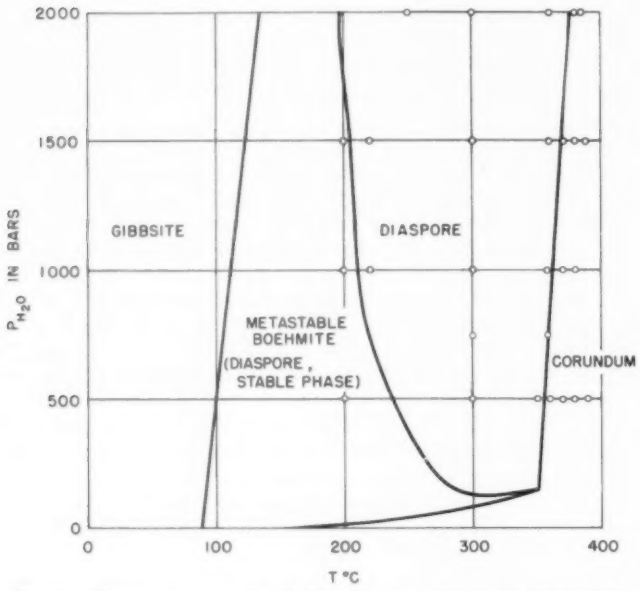

Fig. 2. Phase relations at low H_2O pressures in the system Al_2O_3—H_2O.

number of hydrothermal runs have been made in conventional apparatus. These runs serve to delineate the low pressure part of the phase diagram. Some of the data from low pressure hydrothermal runs are shown in figure 2. The phase boundaries shown in this figure are based partly on the writer's data and partly on the data of Ervin and Osborn. The upper stability limit of gibbsite is sketched in figure 2. This boundary is based on the determined temperature of breakdown of gibbsite to boehmite at 10 kb and on the breakdown temperature at one atmosphere pressure. At one atmosphere pressure, gibbsite breaks down to boehmite at an exceedingly slow rate under appropriate conditions at approximately 90° (R. K. Iler, personal communication). This, of course, does not mean that natural boehmite forms at temperatures above 90° by the breakdown of gibbsite, for at sufficiently low H_2O pressures the breakdown can take place at almost any temperature, although, of course, as temperatures approach ambient temperatures, this transition is very slow. The gibbsite-boehmite boundary as shown in figures 1 and 2 is, of course, not an equilibrium boundary, as boehmite is not a stable phase. Presumably the equilibrium gibbsite-diaspore boundary should be almost coincident with this boundary but should lie at a few degrees lower temperature.

The area labeled "Metastable Boehmite", in figure 2, represents the range of temperatures and pressures in hydrothermal experiments in which metastable boehmite forms. Diaspore is the stable phase in this area. The field of "Metastable Boehmite" as deduced from hydrothermal apparatus is much smaller than the same field shown in figure 1, as determined from the piston-anvil type

apparatus—further evidence that this is a rate boundary and not an equilibrium boundary.

Experimental runs made by the author are shown as circles and were used to delineate the diaspore-corundum boundary and the boundary of the "Metastable Boehmite" field. Many of these runs, particularly the ones in the diaspore field, were maintained at temperature and pressure for as long as three months. Alumina gel was placed in a bomb which was seeded with a small trace of diaspore crystals. The lowest temperature at which the diaspore had appreciably increased in quantity and the crystals increased in size, in three months time, were the two runs shown at 220° at 1000 and 1500 bars H_2O pressure. The lower part of the diaspore field in figure 2 has been sketched in from data by Ervin and Osborn.

The breakdown of diaspore to corundum-plus-water has been delineated by a series of runs, each of approximately one month duration. This curve intersects the vapor pressure curve of liquid water at approximately 150 bars and 350° at which point it must, of necessity, turn and essentially follow along and slightly below the vapor pressure curve of water down to lower temperatures. The curves showing the vapor pressures of the reaction of gibbsite to metastable boehmite-plus-water and of diaspore to corundum-plus-water are both essentially straight and, within the precision of the particular experiments, show no appreciable curvature, unlike vapor pressure curves determined for reactions that take place at higher temperatures. The reason for this is that liquid water, not vapor, is the product of dehydration. The volumes of the liquid water are not particularly sensitive to pressure changes of the order of magnitude of those shown on figure 2. There is, of course, a sharp inflection in the slope of the diaspore to corundum-plus-water curve where this curve intersects the vapor pressure curve of pure water; here, a very large change in Δv takes place.

The high pressure data shown in figure 1, with some analogous curves obtained from studies in the system, SiO_2—Al_2O_3—H_2O, to be published at a later date, are shown in figure 3. Surprisingly, essentially straight lines are obtained at pressures of 4 kilobars and above, when the log of pressure is plotted against temperature. This relation does not hold at low pressures, of course, where water behaves more as an ideal gas. Here the log of the pressure must vary as the reciprocal of the absolute temperature.

The vapor pressure curves for the related hydrates shown in figure 3 are remarkably subparallel to each other, seeming to indicate a rather uniform change in compressibility in these systems as pressure is increased. It will be of interest to learn whether or not other mineral hydrates exhibit vapor pressure curves that are part of this set.

DATA FROM THERMODYNAMIC MEASUREMENTS

Recent thermodynamic data have thrown much light on the possibility of existence of a stable field for boehmite. K. K. Kelley and his colleagues at the Minerals Thermodynamics Experiment Station, particularly E. G. King, have recently made low temperature heat capacity measurements on both boehmite and diaspore. The sample of boehmite used was prepared hydrothermally by the writer for these measurements. The diaspore was from the famous locality

at Chester, Massachusetts and was kindly supplied the writer by C. Frondel. The diaspore was rather impure and, though it was crushed and cleaned as well as possible, the final cleaned product contained 0.4 percent MgO.

The low temperature heat capacity measurements of Kelley and King indicate $S^\circ_{298} = 16.86 \pm .06$ cal/deg/mole for diaspore and $S^\circ_{298} = 23.15 \pm .10$ cal/deg/mole for boehmite ($Al_2O_3 \cdot H_2O$). The correction for the MgO in the diaspore sample ranged from 1.0 to 11.6 percent, depending on the temperature (K. K. Kelley, personal communication, September 15, 1958). Thus the value of Δs for the reaction boehmite to diaspore is $6.29 \pm .16$. The molar volume of diaspore, from x-ray measurements is 35.66 cc, of boehmite 38.65 cc. Thus $\Delta v = 2.99$ cc.

The slope of the equilibrium boundary separating the stable phases diaspore and boehmite should be as follows,

$$\frac{dp}{dt} = \frac{\Delta s}{\Delta v} = \frac{6.29 \pm .16 \text{ cal/deg/mole}}{2.99 \text{ cc}} = \frac{2.63 \text{ deci-joules/deg/mole}}{2.99 \text{ cc}}$$

$$\frac{dp}{dt} = +88 \text{ bar/degree}$$

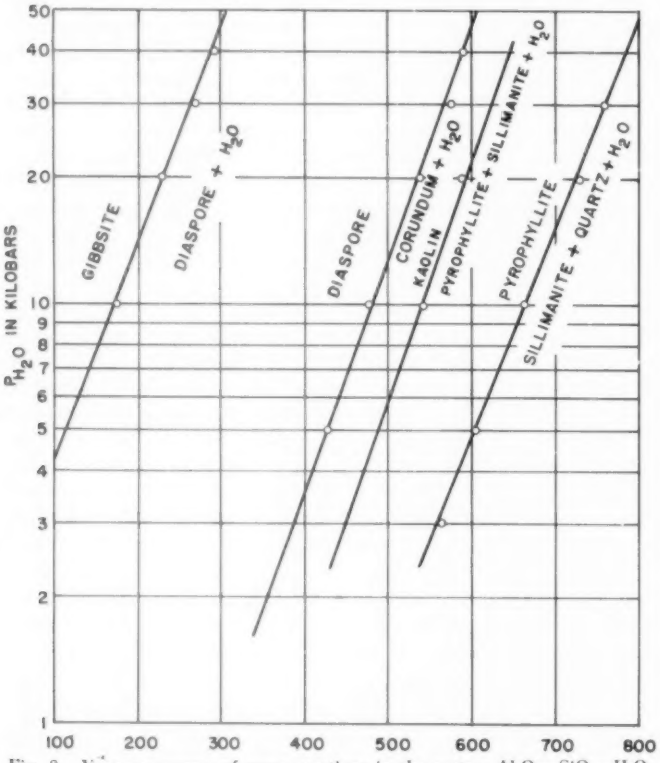

Fig. 3. Vapor pressures of some reactions in the system Al_2O_3—SiO_2—H_2O.

This value, +88 bar/degree, is the largest slope known to the writer for a solid-solid transition among natural minerals and results from the unexpectedly low value for the entropy of diaspore.

It is clear then, within the limitations of the assumption that Δs and Δv are essentially independent of temperature and pressure, that there can be no stability field for boehmite, for the field of boehmite must lie at higher temperatures and lower pressures than the field of diaspore. A line with the slope of +88 bar/degree passed through the highest-temperature lowest-pressure synthesis of diaspore from boehmite, as shown in figure 2 of this paper, or as taken from the data of Ervin and Osborn, shows that corundum lies at higher temperatures and lower pressures and there is no room for a boehmite field.

It thus appears that within the limitations of the thermodynamic and experimental data, boehmite is metastable at all temperatures and pressures and probably always forms metastably in natural deposits. Diaspore is then the stable phase in equilibrium with gibbsite.

An assumption has, of course, been made that the diaspore and boehmite of natural deposits are reasonably pure phases and accept other elements to a very limited extent in their crystal structure. Such analyses of boehmite and diaspore as are available support this view. Though iron is an ubiquitous impurity in "boehmite" samples, it is assumed to be present as the separate phase and not in solid solution relations with boehmite.

DATA FROM THE FIELD

Much field information indicates that diaspore can form on the surface of the earth under essentially no confining pressure, given sufficient time and the right climatic or chemical environment. Indeed all three of the aluminum hydrates—gibbsite, boehmite and diaspore—are present in some bauxite deposits and other bauxite deposits are made up wholly of one or the other of these three phases. The lateritic clays and laterites, which contain all three phases, show geological evidence of having been formed under remarkably similar conditions of pressure and temperature. This strongly suggests that either one or the other of these phases is present metastably. Keller (1952) summarizes this; "Because of the intimate coexistence of diaspore and boehmite it is obvious that the physio-chemical conditions required for their formation in nature cannot be radically different."

There is a strong suggestion that diaspore tends to form most readily from the metastable boehmite in limestone environments. This suggests that $CaCO_3$ saturated solutions are better able to disolve the metastable phase and precipitate the stable phase than are other types of laterite-producing solutions.

Allen (1952) has discussed in detail the mineralogical relationships between the phases gibbsite, boehmite and diaspore. His descriptions emphasize the extremely slow rates of conversion of one phase to another under geological conditions. Boehmite may be associated either with gibbsite, diaspore, or both, and metastable phases may persist over long periods of geologic time. Allen (1935) has described in detail the diaspore clays of Missouri. These clays are found filling sandstone lined depressions and sink holes in rocks of Pennsylvanian age. It seems unlikely that these deposits have formed at pressures of

more than a few atmospheres. Allen presents evidence that oolites of boehmite and diaspore have grown at the expense of clay minerals, and everywhere in the diaspore district, except possibly one mine, diaspore has grown directly from the flint clays without an intervening stage of gibbsite.

Frederickson (1952) is explicit on this point; in discussing the clay deposits of Missouri and Pennsylvania he says, "Boehmite and diaspore coexist in textural relationships that makes it doubtful that these rocks have been subjected to elevated temperatures and pressures."

Most interesting relations between boehmite and diaspore are shown in the Mercer fire clay at North Central Pennsylvania (Bolger and Weitz, 1952). In this clay boehmite and diaspore seem to be contemporaneous and are both growing at the expense of kaolin. Bolger and Weitz, who have examined the mineral relations of the Mercer fire clay deposits in detail, state, "It thus appears quite unlikely that abnormal pressure, abnormal temperature, or epigenic processes, either alone or in combination were responsible for the formation of the earliest diaspore," and conclude that the Mercer diaspore formed at the surface of the earth under conditions of lateritic weathering. Keller (1952) discussing the origin of Missouri high aluminum clays also emphasizes the very intimate mixture of boehmite and diaspore.

It is gratifying that the evidence from the field and from the laboratory are not inconsistent. Diaspore cannot be considered exclusively a high temperature or high pressure phase, and boehmite almost certainly has no field of stability in lateritic or other environments.

ACKNOWLEDGMENTS

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FLOW ORIENTATION OF OLIVINE

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ABSTRACT. Petrofabric analyses of olivine indicatrix orientations in peridotites, basic plutons and intrusive basalts have revealed a common pattern of alignment in which a concentration of α poles lies normal to a girdle containing β and γ in the plane of the megascopic foliation. This fundamental pattern, together with some variations, is ascribed to laminar flow during intrusion. A similar deformation fabric is developed in olivine tectonites where the glide plane (010) of olivine is oriented parallel to the principle s surface.

INTRODUCTION

A survey of literature on olivine-rich rocks from various parts of the world has shown that the orientation diagrams for olivine obtained by petrofabric analysis display a number of consistent features. A table which summarizes many of these olivine fabrics has been published by Ladurner (1954b).

Andreatta (1934) described an olivine rock from the Ortles-Cevedale massif in northern Italy and found from plotting the optical directions of the olivines that α was repeatedly oriented perpendicular to the foliation in the rocks, while β and γ occupied a girdle. Similar features were recorded by Ernst (1935) for olivine-schists from Norway, with α showing a single strong concentration normal to the schistosity, the other directions β and γ lying in a clearly marked girdle. Phillips (1938) examined intrusive fissile dunites from Rum and Skye and in fabric diagrams discovered the usual concentration of α perpendicular to a girdle containing β and γ; in addition, there was a marked and measurable elongation of the olivine crystals in the foliation plane parallel to the $\beta\gamma$ girdle. These results showed quite clearly that there was a strong tendency for olivine in foliated rocks to adopt an orientation which was determined by the common habit of the mineral; that is, the larger faces parallel to (010) coincided with the foliation plane.

Turner (1942) made a detailed study of the fabric of olivine grains in rocks from a number of localities and reached the following conclusions: (a) Olivine crystals suspended in lava do not show a preferred orientation. (b) Gravitative settling of olivine does not produce an orientation. (c) In banded dunites and some Hebridean peridotites the orientation pattern has been produced by the effect of flow in an essentially solid "magma", with interstitial lubrication, on olivine crystals which characteristically are flattened parallel to (010); the typical pattern is a strong concentration of α poles perpendicular to the megascopic foliation and to a girdle containing β and γ. (d) In large dunite intrusions which were emplaced during strong orogenic folding the primary pattern of olivine orientation was determined by flow, but later deformation has possibly modified the original fabric. (e) The criteria for solid deformation are undulose extinction in the olivine grains; the development of translation lamellae perpendicular to γ; the appearance of an α girdle perpendicular to the original flow planes with the formation of a concentration of γ perpendicular to the α girdle and parallel to b axis of the fabric. (f) Where deformation has exceeded the elastic limit of the rock a cataclastic dunite re-

sults with no orientation pattern in the constituent grains.

Turner's (1942) conclusion that the commonest fabric in these rocks, a concentration of α perpendicular to a girdle containing β and γ, was developed by flow has been supported by petrofabric patterns from banded troctolites described by Huang and Merritt (1952) from Oklahoma. The orientation of olivine in these rocks gives a concentration of α which is perpendicular to the banding and to a broad girdle containing β and γ. It is particularly notable that within this girdle there are well defined separate maxima for β and γ, so that neither of these directions alone forms a complete girdle. These troctolite fabrics with mutually perpendicular maxima for α, β and γ, and with the $\beta\gamma$ girdle in the plane of megascopic foliation, are of special interest since a closely similar fabric from a banded dunite was interpreted by Turner (1942) as a result of deformation of the solidified intrusive mass. In the case of the Oklahoma troctolites there is no petrographic evidence to suggest that deformation of the solid rock has taken place, for undulose extinction is rare within the constituent olivine grains while translation lamellae and marginal granulation are absent. The olivine crystals are xenomorphic, but many show a distinct elongation parallel to the $\beta\gamma$ girdle and to the banding in the rock. The fabrics obtained by stereographic analysis appear to have been developed by simple rotation of the crystals to a preferred orientation within planes of laminar flow in a fluid medium.

OLIVINE BASALT FABRICS

When considered in conjunction with Turner's (1942) deductions, Huang and Merritt's (1952) results have indicated quite clearly that the common orientation pattern of olivine in basic and ultrabasic plutonic rocks may be due in many cases to a form of laminar flow in an environment that contained a fluid or other lubricating phase. To test this hypothesis further, orientation diagrams have been prepared for olivine in basic lavas from dykes, volcanic necks and surface flows. The results are interesting, for certain intrusive olivine basalts have yielded patterns which conform closely to the characteristic peridotite orientation where there is a concentration of α perpendicular to a girdle containing β and γ.

Figures 1, 2 and 3 were prepared from olivine basalt (Auckland University thin-section 5176) in a dyke, 10 feet wide, which intrudes a thick compact sequence of tuffs and volcanic breccias at the Bridal Veil falls, on the eastern flanks of Kaikai extinct volcano near Raglan Harbour. A strong concentration of α, with maximum greater than 12 percent, lies normal to a well defined β girdle and to the flow planes in the rock, the maximum deviations in which have been determined from groundmass feldspar lath orientation and entered on figure 1 as broken lines. The β girdle contains a number of distinct small submaxima, none greater than 8 percent, which are fairly evenly distributed throughout, while the angular width of the girdle corresponds closely to the maximum angular deviation of trend in flow planes as measured from the feldspar laths. Poles of γ form an incomplete girdle coplanar with, and occupying a large vector within, the β girdle. In addition, there is a notable tendency for some γ poles to occupy a position normal to the foliation plane

Figs. 1-3. Olivine (50 grains) in basalt dyke (5176). Contours: 2%, 4%, 6%, 8%, 10%, 12% per 1% area. Broken lines show maximum deviations in direction of elognation of groundmass feldspar laths.

Figs. 4-6. Olivine (50 grains) in basalt dyke (5177). Contours: fig. 4, 2%, 4%, 8%, 12%, 16%, 20%; figs. 5-6, 2%, 4%, 6%, 8%.

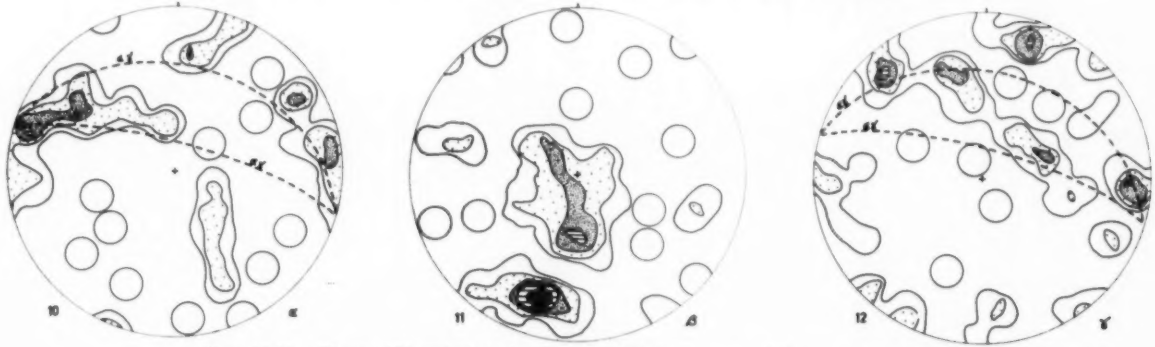
in correspondence with the measured concentration of α in figure 1. The α poles which have been displaced stand out quite clearly in figure 1 where they lie on the trace of the β girdle and thus preserve the flow symmetry of the fabric.

The ability of γ poles to replace α has been noted within strongly oriented petrofabric patterns from other porphyritic olivine basalts and the occurrence of this interchange can be explained satisfactorily by measurements of face size in the crystal habit of the rotated phenocrysts. In the majority of cases, the olivine phenocrysts of these rocks have a characteristic habit flattened parallel to (010), with only minor size development of {Okl} and {001} in the zone [100], and of {110} in the zone [001]. However, in some cases the faces of the form {110} are extraordinarily large and well developed, with corresponding reduction in the size and area of {010}. Such crystals effectively are flattened parallel to (100) and although this face is generally absent, the symmetrical orientation of the large faces (110) and ($1\bar{1}0$) with respect to the flow plane brings γ perpendicular to the general foliation of the rock since γ bisects the low interfacial angle (110) $\wedge$ ($1\bar{1}0$). The easy interchange in position of α and γ poles, depending on the variable crystal habit of the olivine phenocrysts, suggests that $\alpha\gamma$ girdles may be interpreted as the result of alternate preferred orientation positions in a fluid medium. Furthermore, where the normal to such a girdle is a β-maximum lying in the foliation, then this direction could be regarded as a linear element in the flow plane indicating the direction of movement.

Olivine basalt (5177) from a similar dyke at Kirikiripu volcano, Raglan, has almost an identical pattern of olivine orientation (figs. 4, 5, 6). A maximum concentration for α is perpendicular to the foliation expressed by the small feldspar laths, while β and γ together form a clear girdle. The submaximum for γ is strong on the 8 percent contour and it may be significant that small γ submaxima away from the $\beta\gamma$ girdle can be aligned on the trace of a great circle (fig. 6) which coincides with the position of an incipient $\alpha\gamma$ girdle. Although a complementary 8 percent concentration for β is reflected at the south end of the $\beta\gamma$ girdle, other β submaxima of the same value are spread throughout the foliation plane.

Other dykes of similar rock from various localities in New Zealand have been examined for fabric patterns in olivine, but none have yielded orientations as clear as the two described above. In the case of the Ngatutura basalt intrusion at Port Waikato and Marehemu volcanic neck at Kaipara the imperfections in pattern possibly have been caused by physical conditions associated with the vertical attitude and large size of the intrusive bodies, up to 300 yards in width at Ngatutura. Lava from the neck of the Marehemu vent, Kaipara Harbour (114) provides a possible example of this type (figs. 7, 8, 9) where the broader areas of concentration for α, β and γ occupy positions which are approximately orthogonal within the fabric. On a combined plot β and γ form a distinct, but not sharp, girdle perpendicular to which lies the trace of a weakly defined α girdle containing a maximum concentration area greater than 10 percent normal to the plane of β and γ. Furthermore, there is a marked coplanar arrangement of γ with α in additional $\alpha\gamma$ girdle normal to

Figs. 7-9. Olivine (50 grains) in basalt neck (114). Contours: 2%, 4%, 6%, 8%, 10%.



Figs. 10-12. Olivine (50 grains) in basalt dyke (1446). Contours: 2%, 4%, 6%, 8%, 10%.

the trace of the $\beta\gamma$ plane. Although the distribution of the main maxima in the diagrams points to an exceptionally strong orientation of the olivine indicatrices, the fabric as a whole is somewhat diffuse and contains a number of well-dispersed poles. Possible explanations for this condition may be found in two processes, perhaps acting concurrently: (a) The intrusive basalt ceased to flow, but cooled slowly and allowed time for gravitational movements of the olivine crystals, thus modifying a highly developed flow orientation; (b) A fast rate of lava movement through the volcanic throat produced a flow pattern, approaching the ideal orthogonal orientation for α, β and γ, which was marred by relaxation and expansion of the lava body in a zone of reduced hydrostatic pressures in the upper part of the lava column.

An example of dilation causing imperfections in the orientation pattern has been found in olivine basalt (1446) which forms a dyke averaging 4 feet in width in loosely compacted fire-fountaining cinders in the grounds of the Teachers Training College, Auckland. This intrusion was originally described by Bartrum (1928). The weak pattern recorded in the orientation of the olivine (figs. 10, 11, 12) is unusual in several respects. Poles of α and γ show no major concentrations above 8 percent within the projection, but it is possible to recognize a wide poorly defined $\alpha\gamma$ girdle which has an angular width of about 40° (figs. 10 and 12). On the other hand, the two broad β maxima are well defined (fig. 11), but they show no definite elongation in the manner of a girdle and lie 40° apart across their main concentrations in sympathy with the width of the $\alpha\gamma$ girdle. In a preceding section the conclusion was reached that a β-maximum normal to a girdle containing α and γ could represent the direction of flow within laminar planes of movement. The weak pattern of figures 10, 11 and 12 must then be interpreted as the result of inefficient flow movement within the dyke, together with buckling of the laminar flow planes. Field evidence supports this view, for the main dyke thickens and thins in an irregular fashion, while small subsidiary offshoots penetrate the enclosing scoria. The inference drawn from this example is that modifications in the flow orientation of the olivine have been caused by a distinct reduction in the containing hydrostatic conditions when the lava entered the porous unconsolidated scoria layer. Lowered hydrostatic pressure allowed the dyke to bulge and dilate, with flexuring of an underdeveloped flow pattern. Although the dyke is aligned vertically, the narrow width and the environment of emplacement must have induced quick cooling in the lava sheet before the weak flow orientation of the phenocrystic olivine was disturbed by gravitational sinking.

Petrofabric analyses have also been made of olivine indicatrix orientations in selected platy basalt flows near Auckland. A flow at Wiri (4649) contains abundant euhedral olivine phenocrysts, but the orientation diagrams (figs. 13, 14, 15) lack a definite flow symmetry pattern and minor maxima for α, β and γ are spread throughout the projection. In a similar fashion, a flow (3560) at the foot of Mt. Eden has given a dispersed pattern of indicatrix axes, with only a slight tendency for grouping in specific vectors within the plot (figs. 16, 17, 18). Fuller (1939) has commented upon the variable rate of flowage in basalt flows, due to temporary ponding and changes in rate of

extrusion, so that phenocrystic olivine consistently tends to settle downwards. The absence of a clear flow orientation for olivine in such a setting may prove to be the general rule.

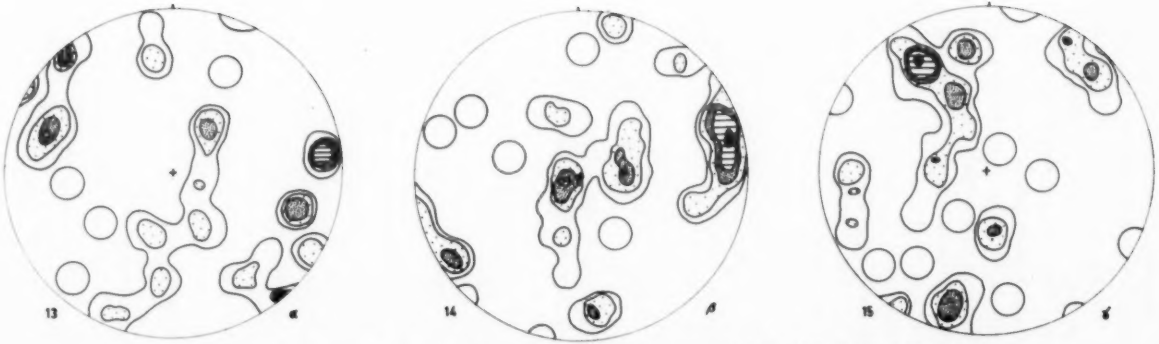
DISCUSSION

The strong similarity between the olivine fabrics in peridotites, banded basic plutonics and intrusive basalts suggests that in all cases there is a common preferred orientation which has been developed by the movement of crystals in a fluid medium. The commonest orientation shows a concentration of α perpendicular to a girdle containing β and γ, but variations of this pattern may occur where the crystal form of the olivine grains provides large faces other than (010) which can lie in or nearly parallel to the foliation plane.

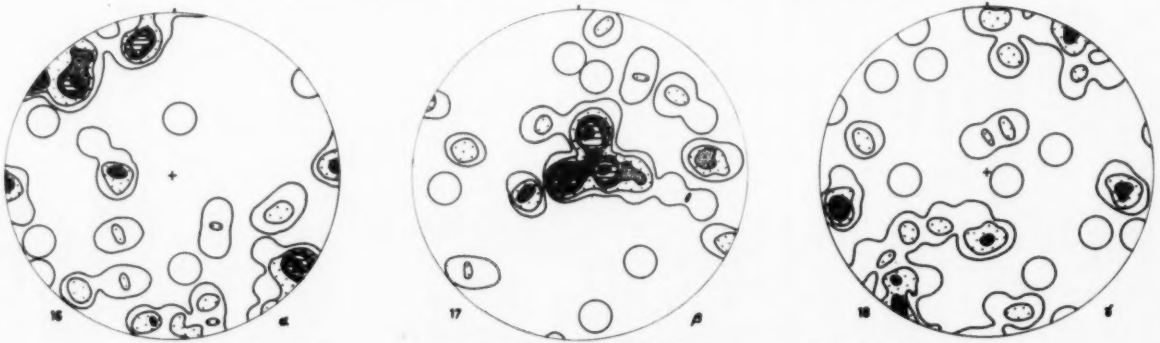
In the case of the ultrabasic rocks, as pointed out by Turner (1942), interstitial lubrication between the grains must have allowed the crystal mush to flow during intrusion. There seems to be no necessity to postulate post-solidification deformation in order to explain certain of the orientation patterns described by Turner (1942) and Fairbairn (1949) as tectonite fabrics, since some of these patterns have now been duplicated from rocks which undoubtedly were highly liquid at the time of emplacement and have not since suffered appreciable orogenic distortion. Included here are fabrics where α, β and γ form mutually perpendicular maxima; the presence of a strong maximum for γ or an incipient α girdle; and the development of a girdle of α and γ with a clear β-maximum lying normal to the plane of the girdle and parallel to the foliation.

There is a strong possibility that in many peridotite orientation patterns the controlling factor in alignment of olivine grains has been crystal form. Measurements of tabularity and elongation of these grains (e.g. Ladurner, 1954b) invariably are made normal to the optical indicatrix axes and expressed in terms of the pinacoids (100), (010) and (001), but such methods cannot reveal the presence of prisms and domes, which from the earliest seed crystal stage onwards, may influence the preferred orientation (Knopf and Ingerson, 1938, p. 134-135). In this connection it is interesting to note the fabrics recorded by Ladurner (1954a, 1954b) in a banded chromite dunite from Turkey. Ladurner prepared separate petrofabric diagrams for olivine grains in the chromite-rich layers and for olivine in the olivine-rich layers and found that the two orientation patterns were quite dissimilar, a difference in fabric which should not have occurred if the olivine alignment had been caused by solid deformation. However, Ladurner (1954b) has suggested that the olivine-rich pattern was due to deformation twinning on (110) since duplicated maxima for γ appear in the diagram across an arc of 47° which is, in fact, close to the interfacial angle between the possible large faces (110) and (1$\bar{1}$0), either of which could lie in the foliation plane. At the same time, Ladurner (1954b, p. 33) has pointed out for the fabric of another sample from Nonsberg that there may be a relation between grain shape and grain orientation such as could be caused by the presence of faces of the form {120}.

The β-maximum pattern, like those described by Huang and Merritt (1952) and in figure 11 of this paper, where poles to (010) tend to occupy



Figs. 13-15. Olivine (45 grains) in basalt flow (4659). Contours: 2.2%, 4.4%, 6.6%, 8.8%, 11.1%.



Figs. 16-18. Olivine (50 grains) in basalt flow (3560). Contours: 2%, 4%, 6%, 8%, 10%.

girdles, has been regarded by Fairbairn (1949, p. 146-147) as evidence of a process of solid deformation which possibly was responsible for the formation of translation lamellae parallel to (100), that is, intergranular rotation about [001] developing strain which is relieved by flexure gliding on (010) parallel to [100]. The occurrence of a β-maximum pattern in olivines in a fluid setting does not entirely disprove Fairbairn's theory, but it does indicate that strain causing the development of shadow bands and lamellae parallel to (100) may be due to some process other than the mechanical stress of peridotite solid intrusion, or of an inferred post-intrusion deformation. In addition, the recognition in basaltic rocks of an olivine orientation where poles to (001) form a maximum within the foliation plane, and normal to a girdle of α and γ, is contrary to Fairbairn's (1949, p. 148) suggestion that this pattern may be characteristic of a tectonite solid flow fabric. Orthogonal maxima for α, β and γ, with α perpendicular to the foliation, may be the ultimate streamline alignment of the common tabular crystal habit of olivine under conditions of strong laminar flow in a lubricated environment.

A high viscosity during intrusion undoubtedly is an important factor in the preservation of flow orientation of olivine, with a maximum for α normal to a $\beta\gamma$ girdle, in the type of gabbro described by Huang and Merritt (1952). Elsewhere in the same locality Huang and Merritt have recorded similar troctolite bodies in which the olivine crystals have undergone gravitative sedimentation within the parent melt and display no preferred orientation. Preservation of the olivine flow fabric in a cooling dyke or volcanic neck probably depends upon a number of physical factors which in part are controlled by the environment of intrusion. Optimum conditions for the development and maintenance of the flow pattern within a dyke appear to be: narrow width, steady rate of intrusion, retention within walls of rigid country rock in order to maintain hydrostatic pressure and prevent dilation, and fairly quick cooling with concurrent rapid increase in viscosity. As a result, not all dyke forms reveal the typical flow orientation of olivine phenocrysts and in the present study 10 intrusions yielded only 5 recognizable patterns.

The common orientation of olivine, a concentration of α perpendicular to a girdle containing β and γ, clearly indicates that a planar element is dominant during flow, for α is the normal to the form {010} which usually possesses by far the largest faces in the flattened habit of olivine. The importance of laminar flow in dyke formation has been emphasized by Spry (1953) who carefully deduced that turbulence could occur only in the event of low viscosity due to superheating or high volatile content, and that in most dykes movement will be of the laminar type. The persistence of the planar flow pattern recorded in this paper from a variety of intrusive settings is in accordance with Spry's (1953) results.

The foregoing discussion has emphasized the importance of flow mechanisms in the orientation of olivine in peridotites, gabbros and basalts. Nevertheless, it must be remembered that Tilley (1947) has recorded from St. Paul's Rocks a fabric, with α perpendicular to striping or "fluxion" banding, in peridotites which undoubtedly have suffered solid deformation. The oriented olivine lattices were measured by Tilley in the finely granular cata-

clastic groundmass of dunite-mylonites and this textural setting offers clear proof of orientation by dynamic forces, in contrast with the apparently non-granulated α-maximum fabrics described by Turner (1942) and in this paper. Other fabrics of the same type, which appear to be similar in origin, are those described by Andreatta (1934, 1946) from tectonites in Northern Italy, the Vanelven olivine-schists examined by Ernst (1935), and banded chromite dunites from Greece (Paulitsch, 1953) which have been deformed by close-spaced shear planes lying at about 50° to the chromite banding. In all these examples a relatively simple pattern of orientation has been determined, with (010) parallel to the foliation, except in the Greek dunites where in crystals tabular parallel to (001) the corresponding perpendicular [001] is oriented normal to the banding and to the fissility.

It seems likely that two entirely distinct processes of alignment, fluid flow and solid deformation, are capable of producing closely similar patterns of orientation in olivine rocks. The strong affinity shown by these highly developed fabrics in the two different settings may be due to the fact that the commonly present large side-pinacoid (010) tends to lie in planes of laminar flow foliation in the fluid environment, while the same face acts as the glide plane which is oriented parallel to the principal *s* surface foliation during deformation and flexure gliding in olivine tectonites (Turner, 1942; Chudoba and Frechen, 1950). The fabrics of some deformed peridotites may thus be only slightly modified versions of the olivine orientation patterns previously developed during semi-solid intrusion, a possibility pointed out earlier by Turner (1942).

ACKNOWLEDGMENTS

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CHRONOLOGY OF MAJOR METAMORPHIC EVENTS IN THE SOUTHEASTERN UNITED STATES*

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ABSTRACT. Potassium-argon and rubidium-strontium age measurements have been made on a variety of granites, pegmatites, gneisses and schists which comprise the plutonic-metamorphic complex of the Piedmont and Blue Ridge of the southeastern United States. Large portions of the area appear to have been metamorphosed initially approximately at the same time as the Grenville Province, i.e., about 900 to 1100 m.y. ago. Superimposed on this older metamorphic province was a major orogenic event culminating at about 350 m.y. with widespread recrystallization of existing rocks and intrusion of pegmatites in the Spruce Pine, Franklin-Sylva and Bryson City Districts, and granites in the Virginia and North Carolina Piedmont. There is strong evidence of an additional metamorphic epoch between 350 and 1000 m.y., but its effects have been largely obliterated by the 350 m.y. event.

In western North Carolina a transition of apparent ages from 355 to 890 m.y. occurs in the same rock unit (Cranberry gneiss) over a distance of about 10 miles across the strike of the border of the 350 m.y. event.

In the southeastern Piedmont of Georgia and South Carolina a younger metamorphic event or events can be detected producing rocks of apparent age ranging from 230 to 310 m.y. The time of these orogenies is compared with those in the Central and Northern Appalachians.

Evidence is accumulating that the Holmes' time scale will have to be considerably lengthened.

INTRODUCTION

The absolute age relationships of the plutonic-metamorphic complex of the Blue Ridge and Piedmont of southeastern United States have been essentially unknown. The most certain relative ages were that late Precambrian sediments (the Ocoee series) rested unconformably on a basement complex along the western border of the Blue Ridge and that lightly metamorphosed Paleozoic(?) sedimentary and volcanic rocks covered broad belts of a metamorphic basement on the eastern Piedmont (Rodgers, 1956; King, Hadley, Neuman and Hamilton, 1958). In the Spruce Pine District of western North Carolina it was shown that there were at least two distinct periods of regional metamorphism separated by a period of dike intrusion (Kulp and Poldervaart, 1956). The latest of these was accompanied by the emplacement of the pegmatite swarm in Lower Paleozoic time. No information was available on the time of the first metamorphism which initially produced the basement complex, although reconnaissance work by Carr and Kulp (1957) using the potassium-argon method on bulk rock samples had shown that the basement to the west of the Spruce Pine District was clearly older than the pegmatites. In view of the thick sequence of Precambrian sediments (Ocoee) overlying this basement, it was suspected to be Grenville in age. These basement units or lithologic facsimiles can be traced eastward nearly to the Cretaceous overlap so that a clear determination of the time of initial metamorphism at any one point would probably define the time of formation of the primary basement of the Southern Appalachians and Piedmont.

This investigation was directed to the definition of the time of the major

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metamorphic events in this region. The establishment of the potassium-argon (Wasserburg and Hayden, 1955; Wetherill, Aldrich and Davis, 1955; Reynolds, 1956; and Carr and Kulp, 1957) and rubidium-strontium (Davis and Aldrich, 1953; Aldrich, Wetherill and Davis, 1956; and Aldrich, Wetherill, Tilton and Davis, 1956) methods of age determination on micas of metamorphic and igneous rocks made such a study feasible. A suite of samples was collected which represented the major rock units over the region. More detailed collections were made in western North Carolina where preliminary age determination work had indicated that the first metamorphic event which produced the present basement rocks might be revealed. The geographic and petrographic descriptions of these samples are given in the Appendix.

The authors wish to acknowledge the generous support of the Research Division of the Atomic Energy Commission which has made this investigation possible. They are indebted to their colleagues P. E. Damon, P. W. Gast, P. M. Hurley, John Rodgers, Arie Poldervaart, G. R. Tilton and K. K. Turekian for helpful discussions and suggestions. In the experimental work we were assisted by F. Bazan, C. Giffin, and G. Hall. The typing and drafting were done by M. Rippey and J. Brokaw.

EXPERIMENTAL MEASUREMENTS

The rock samples were crushed and the mica separated by a combination of heavy liquid, mechanical and magnetic methods. The rubidium and strontium contents were determined by the method of isotope dilution as employed in this laboratory (Gast, Kulp and Long, 1958). For relatively young samples or for samples with a low Rb/common Sr ratio, the uncertainty in the Sr^{87} correction is the dominant error in the analysis. The mass spectrometer errors are on the order of 1 percent for the Sr^{87} abundance. The errors given on the ages (table 1) represent the probable standard error from the analytical procedures. The argon content of the minerals was also determined by isotope dilution using direct fusion techniques as described by Carr and Kulp (1957) and as modified by Damon and Kulp (1957). This measurement is reproducible to better than 3 percent. The potassium analyses were done by Ledoux and Company and by the University of Rinnesota Rock Analysis Laboratory. The results from the two laboratories on split samples showed a standard deviation of about 2 percent. The standard deviation of the A^{40}/K^{40} ratio is about 4 percent. The analytical data are given in table 2 along with three determinations on Spruce Pine pegmatites reported by Carr and Kulp (1957).

The half life for Rb^{87} was taken as 5.0×10^{10} years (Aldrich, Wetherill, Tilton and Davis, 1956). The gamma decay rate for K^{40} as determined by Wetherill (1957) was adopted yielding $\lambda_e = 0.583 \times 10^{-10}\text{yr.}^{-1}$. The beta decay constant of K^{40} is less certain. A value of $\lambda_\beta = 4.72 \times 10^{-10}\text{yr.}^{-1}$ was adopted. For the relatively late Precambrian and Paleozoic rocks considered in this paper the calculated ages are not very sensitive to the choice of this constant. All ages from other laboratories included in the discussion have been recalculated with these decay constants so that they are strictly comparable.

The decay constants for K^{40} or Rb^{87} probably permit absolute age calculation to within about $\pm$ 5 percent. Relative ages, however, can be established

TABLE 1

Rubidium-Strontium Age Measurements in the Plutonic-Metamorphic Province of Tennessee—North Carolina

Sample No. and Mineral	Sample Weight	Sr (normal)	Sr^{87*}	$\frac{Sr^{87*}}{\text{Total } Sr^{87}}$	Rb	$\frac{Sr^{87*}}{Rb^{87}}$	Apparent Age	Rock Unit	Location
	gm	ppm	ppm		ppm		m.y.		
L-141B	1.5869	15.1	0.594	0.36	457	0.00459 ± 0.00018	330 ± 13	Carolina gneiss	Rte. 19 near Stocksville, N.C.
L-22B	2.1279	25.7	1.52	0.46	880	0.00611 ± 0.00024	439 ± 17	Beech granite	Rte. 19E near Hopson, Tenn.
L-22F	1.0259	64.1	0.485	0.098	215	0.00796 ± 0.00040	572 ± 29	Same location as above	
L-332B	1.3991	11.8	2.03	0.71	717	0.0100 ± 0.0003	719 ± 25	Cranberry gneiss	Near Rte. 19E bet. Hopson and Hampton, Tenn.
L-149B	1.8673	9.4	3.04	0.82	865	0.0124 ± 0.0004	892 ± 30	Cranberry gneiss	Pardee Point, Tenn.
L-140B	0.8653	6.2	0.480	0.52	368	0.00462 ± 0.00021	333 ± 15	Carolina gneiss	Rte. 23 at Gay, N. C.
L-133B	0.7625	11.1	0.801	0.44	620	0.00457 ± 0.00018	329 ± 13	Ocoee formation	Rte. 64 near Ducktown, Tenn.

* Indicates radiogenic component
B Biotite
F K Feldspar

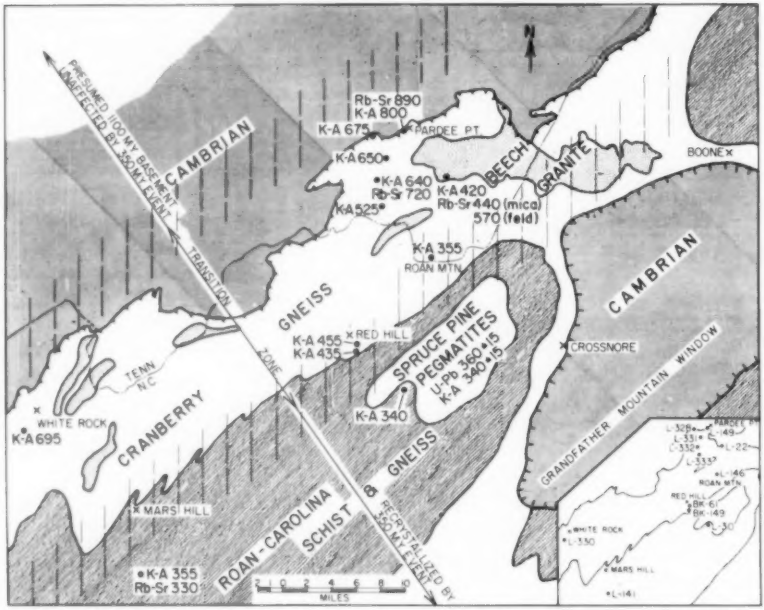

Fig. 1. Isotopic ages and regional geology in the vicinity of the Spruce Pine Pegmatite District showing parts of western North Carolina and eastern Tennessee.

to within about 3 percent, which in the age range of 350 m.y. is on the order of plus or minus 10 m.y. for the standard error.

THE TRANSITION ZONE WEST OF THE SPRUCE PINE DISTRICT OF WESTERN NORTH CAROLINA

The Spruce Pine District is located on the Blue Ridge Plateau in western North Carolina (figs. 1 and 2). The rocks in the area consist of mica and hornblende gneisses and schists, quartzofeldspathic gneisses, ultramafic rocks, pegmatite, alaskite, granite, metadiabase and metagabbro and younger unmetamorphosed diabase. These basement rocks include a vast metasedimentary sequence, the lowermost unit being the Cranberry gneiss which is overlain by the Roan-Carolina gneiss-schist series as shown by Eckelmann and Kulp (1956) (see fig. 1). In the immediate vicinity of Spruce Pine, covering an area of 300 square miles, a large number of pegmatites has been emplaced. To the northwest in the vicinity of Roan Mountain occurs a swarm of basaltic dikes striking northeasterly, which has been metamorphosed but not plastically deformed. In the Cranberry unit to the northwest and west of the Spruce Pine pegmatite district, are many intrusive granites with possible secondary foliation, such as the Beech granite (fig. 1). These granites appear to have been emplaced in conjunction with the first major regional metamorphism, but the basaltic dikes were intruded subsequently in cold fractured rocks prior to the second major metamorphism, since they are in turn cut by pegmatite dikelets

considered to be of the same age as the main Spruce Pine pegmatites. To the northeast is the Grandfather Mountain Window in which Cambrian rocks are exposed as a result of late-Paleozoic thrusting and subsequent erosion.

The ages obtained in this district, rounded off to the nearest 5 million years, are shown in figure 1. For the Spruce Pine pegmatites, existing Rb-Sr and U-Pb ages average about 360 m.y. (Eckelmann and Kulp, 1957; Aldrich, Wetherill, Davis and Tilton, 1958) with the average K-A age being slightly lower. The date of the height of the last major regional metamorphism in this area probably occurred about 350 m.y. ago. Southwest along strike, 30 miles distant, a similar age is obtained in the basement rocks. However, proceeding north-northwest toward Pardee Point (perpendicular to the main structural trend) both K-A and Rb-Sr ages progressively increase. This represents a transition zone where the older basement was only partially recrystallized by the 350 m.y. metamorphic event. At the most northwesterly point the apparent age by the Rb-Sr method increases to 890 m.y.; southwest along strike at White Rock, an apparent K-A age as high as 695 m.y. is reached before the Paleozoic cover prevents further sampling.

Unfortunately, no basement rocks are exposed at the surface northwest of Pardee Point. Northeast along strike but west of the Cambrian contact, such rocks occur in the vicinity of Mountain City as thrust slices, but they are so badly sheared that proper samples could not be procured. In this area unsheared basement rocks were presumably unaffected by the later metamorphism to the east.

In view of apparent ages of 850 to 900 m.y. obtained by the Rb-Sr and K-A methods on micas by Long, Cobb and Kulp (1959) in the New Jersey-New York Highlands in a metasedimentary complex that was probably formed initially 1100 m.y. ago according to a concordant zircon age, and of similar ages near the Shenandoah Valley (Tilton, Wetherill, Davis and Hopson, 1958), it appears that the original metamorphic complex in the Southern Appalachians also formed about 1100 m.y. ago, i.e., at roughly the same time as the Grenville Province of the Canadian Shield (Eckelmann and Kulp, 1957; Aldrich, Wetherill, Davis and Tilton, 1958).

The inferred boundaries of the transition zone if an 1100 m.y. age is assumed are indicated in figure 1. This transition zone may have been somewhat wider originally and has been narrowed by thrusting.

Parenthetically it may be noted that the K-A ages on the mica across the transition zone agree broadly with the Rb-Sr ages (L-141, L-22, L-332, L-149) suggesting that unless the temperature is high enough for appreciable diffusion of the potassium ions in the interlayer position in the mica, neither the argon nor the radiogenic strontium can be lost. The results on the Beech granite, however, indicate that under the same conditions effective diffusion occurs more readily in mica than in feldspar, for here the Rb-Sr and K-A ages on the mica are approximately 130 m.y. younger than the apparent Rb-Sr age on the feldspar. Tilton, Wetherill, Davis and Hopson (1958) reported an even more extreme case where the Rb-Sr age of the feldspar in the Baltimore gneiss was 1150 m.y. in agreement with the zircon ages, whereas the associated micas gave Rb-Sr and K-A ages of about 320 m.y.

G. R. Tilton (private communication) has investigated accessory zircons in the rocks of the Spruce Pine District and finds Pb^{207}/Pb^{206} ages as high as 800 m.y. as far east as Crossnore (fig. 1). This again suggests that the zircons of the basement rock were more resistant to the loss of daughter isotopes than the micas.

The results on the biotite from the diabase dikes near Bakersville, N. C. are particularly interesting as they suggest the existence of a third metamorphic cycle that occurred prior to 455 m.y. ago but subsequent to the presumed 1100 m.y. event. These dikes have sharp contacts and chilled margins and intrude brecciated Cranberry gneiss (Kulp and Poldervaart, 1956). There is some evidence from remnant amygduloidal texture (Wilcox and Poldervaart, 1958) that they were intruded at shallow depths. Thus these dikes clearly came in after the initial regional metamorphism that formed the basement, probably around 1100 m.y. ago, yet they are now metamorphosed (Wilcox and Poldervaart, 1958) and in places biotite has been developed. This biotite, however, could not have been formed during the 350 m.y. event since the A^{40}/K^{40} gives an apparent age of at least 455 m.y. The close proximity of these dikes to the 350 m.y. zone of complete reconstitution of the micas makes it probable that the 455 m.y. age is minimum for this event and is not to be construed as the actual date.

CROSS SECTION THROUGH THE PLUTONIC-METAMORPHIC PROVINCE

Figure 2 is a generalized geologic map of the Southeastern U. S. on which the age measurements are indicated. Southwest from the Spruce Pine District along the major structural trend are two other pegmatite districts, near Franklin, and near Bryson City. The K-A ages from muscovites from both districts agree with the Spruce Pine age at about 350 m.y. (The preliminary age of 540 m.y. on a Bryson City biotite, reported by Carr and Kulp (1957), is now known to be incorrect because of an erroneous potassium analysis.) A sample of biotite (L-140) from the Carolina gneiss country rock within the Franklin-Sylva District gave an apparent K-A age of 440 m.y. Since the Carolina gneiss was part of the primary basement, this age indicates that the radiogenic argon in this biotite was not all lost during the 350 m.y. event. The apparent Rb-Sr age broadly agrees with the pegmatite age possibly suggesting that this isotopic ratio was more sensitive to the effects of the 350 m.y. event. The older K-A age is not unexpected since Olson (1946) has shown that the Franklin-Sylva pegmatites are identical mineralogically with those at Spruce Pine but have an entirely different structural habit, i.e., they were intruded along joints *across* the structure, whereas at Spruce Pine they are almost all conformable and were intruded into obviously plastic country rock. It appears therefore that recrystallization at Franklin-Sylva was not as extensive as at Spruce Pine, and hence the Carolina gneiss at Franklin-Sylva might be expected to retain a higher apparent age.

Along strike to the southwest, muscovite and biotite separated from a sample of Talledega schist (L-132) from McCaysville, Georgia, record the 350 m.y. event. The rocks called Talledega here probably represent Lower Paleozoic sediments but this is not certain.

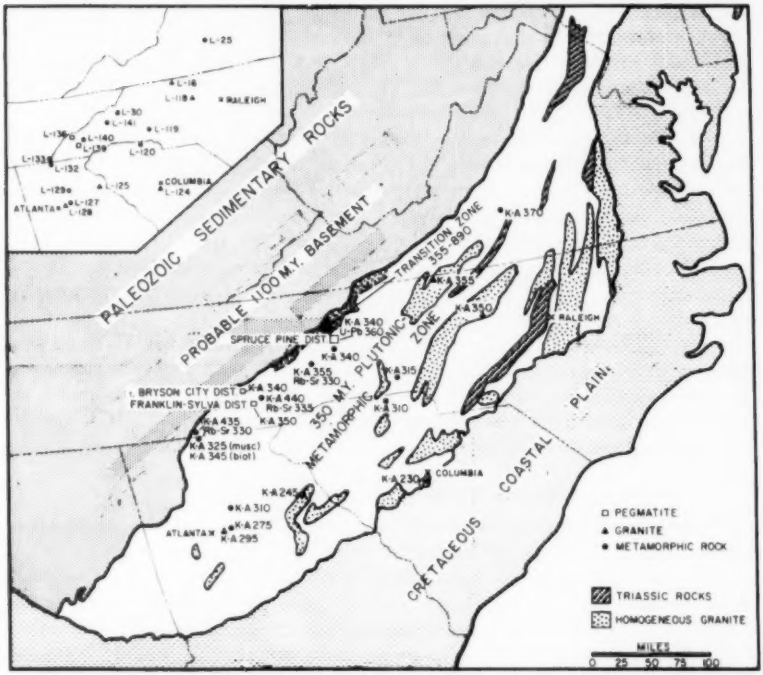

Fig. 2. Generalized geologic map of the Southeastern U. S. on which the isotopic age measurements are indicated.

In the vicinity of Ducktown, Tenn., the Ocoee series consists of low-grade phyllitic mica schist. Its apparent K-A age of 435 m.y. (L-133) virtually establishes a distinct regional metamorphism prior to the 350 m.y. event but subsequent to that of the basement, on which the Ocoee series lies unconformably. Petrographic evidence rules against a possible detrital origin of the Ocoee biotite. Whether the metamorphism was Precambrian or early Paleozoic cannot be decided from this datum. The low Rb-Sr age of the Ocoee biotite and the proximity of this locality to McCaysville where 325 to 345 m.y. K-A ages were obtained shows that the Ocoee rock was affected by the 350 m.y. event and therefore the 435 m.y. K-A age may be construed as a minimum age only.

In the Piedmont east of the Spruce Pine pegmatite district, large bodies of undeformed granite intrude the basement complex. These granites appear to have been emplaced at approximately the same time as the pegmatites in western North Carolina from the K-A age data shown in figure 2. If the granites are truly Carboniferous, as has been inferred by some geologists, these ages add to the accumulating evidence for a lengthened Paleozoic time scale.

THE SOUTHERN APPALACHIAN REGION

Figure 2 shows that the 350 m.y. date is the dominant one throughout

the southeastern Blue Ridge and northern Piedmont. In the southeastern Piedmont, however, there is evidence of a distinctly younger event or events. The K-A dates in South Carolina and Georgia obtained in the present study and the Rb-Sr dates on the same rock units reported by the M.I.T. group (Pinson, Fairbairn, Hurley, Herzog and Cormier, 1957) are given in table 3.

The wide range in ages from the Stone Mtn. area, Georgia, to the South Carolina-North Carolina border indicates considerable complexity in the Paleozoic metamorphic history. A possible hypothesis is that the entire southeastern Piedmont participated in the 350 m.y. event but that subsequently another metamorphic pulse centering to the southeast of the exposed basement caused variable alterations in the isotopic ratios. The 310 m.y. date on the ancient Carolina gneiss (L-129) at the Buford dam site probably indicates the presence of another transition zone.

Grunenfelder and Silver (1958) report nearly concordant U-Pb isotopic ages on zircons separated from the Elberton granite and Lithonia gneiss suggesting crystallization at 450 m.y. or earlier. This is not inconsistent with the evidence deduced for an earlier event in the southeast from the K-A ages on the samples from the Bakersville dikes and the Ocoee rocks discussed above.

A COMPREHENSIVE APPALACHIAN PICTURE

The ages obtained from various laboratories throughout the Appalachian region from Nova Scotia to Georgia are shown in figure 3.

Around Baltimore the rocks have been completely reworked by an event that appears close to 320 m.y. Here A^{40}/K^{40} ratios on micas from the Cockeysville marble, the Setters quartzite, the Precambrian Baltimore gneiss and intrusive pegmatites give a range of values from 300 to 360 m.y. (Wasserburg, Pettijohn and Lipson, 1957), somewhat outside the experimental error. Tilton, Wetherill, Davis and Hopson (1958), working in the same area, confirmed these results using the Rb-Sr method on micas. From the Baltimore gneiss, they obtained nearly concordant U-Pb ages on two zircon samples, which pointed to a true age of about 1100 m.y., and a Rb-Sr age on a feldspar, which gave 1200 $\pm$ 200 m.y. This establishes the probable age of the primary formation of the basement in this area as the same as that in the Grenville Province. The major Paleozoic metamorphism appears to have occurred at a somewhat later time here than in the New York and North Carolina areas, i.e., at about 320 m.y. ago. Near Radnor, Pa., where it is possible to get much farther northwest across the strike of 320 m.y. metamorphism than at Baltimore, an age of 395 m.y. was obtained in the present study on a mica in the Baltimore gneiss (L-41) by the K-A method (table 2). Davis, Wetherill, Tilton and Hopson (1958) also report older apparent ages in this vicinity.

In the New York City area (Long and Kulp, 1958), a number of K-A measurements on the Manhattan Prong rocks give ages in the 360 to 370 m.y. range, whereas a sample from the Fordham formation gives a K-A age of 440 m.y. indicating that the rocks of the New York City group were also produced initially by an older metamorphism. To the north in the New York Highlands, a rapid transition from 360 to 800 m.y. in apparent age is observed much as in western North Carolina. A concordant U-Pb age of 1150 m.y. on

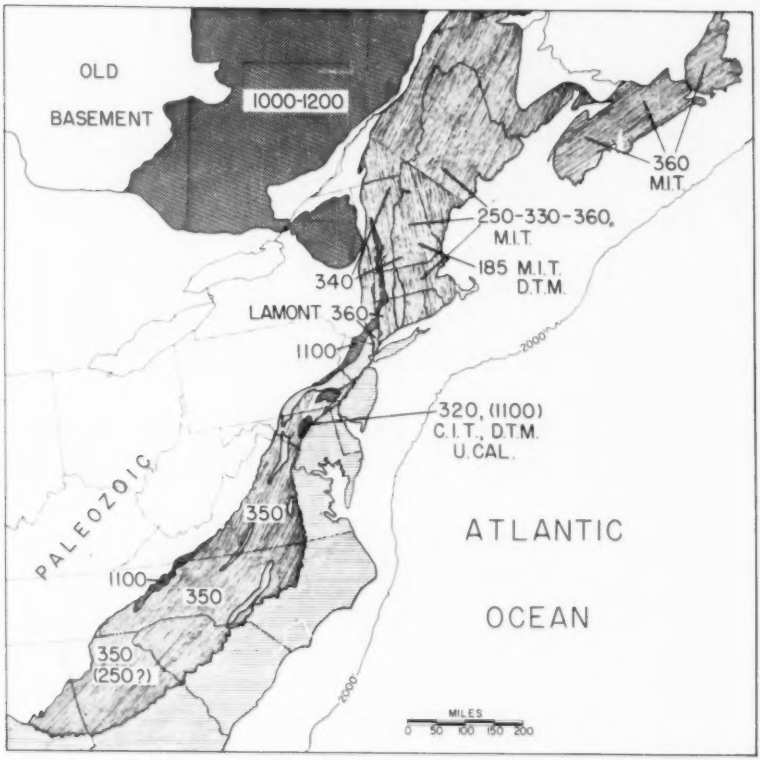

Fig. 3. The probable age relations and regional geology of the basement of the eastern part of North America.

zircon from the Canada Hill gneiss at Bear Mt. (Tilton, Wetherill, Davis and Hopson, 1958) further establishes the rough simultaneity of the formation of the Appalachian-Grenville basement.

In the Northern Appalachians K-A dates on micas in western New Hampshire (Damon and Kulp, 1957) and in the Westmore granite in Vermont (table 2) gave 325 and 345 m.y., respectively. Many measurements by the group at the Massachusetts Institute of Technology on igneous and metamorphic rocks from Massachusetts, Maine and Nova Scotia, (Hurley, Fairbairn and Pinson, 1958) suggest a 360 m.y. age for the major event in the Maritime Province becoming slightly younger to the southwest. This major event was followed by other pulses of decreasing importance in the Northern Appalachians, the more definite ones appearing at about 310 in Nova Scotia, northern New Hampshire, and Maine; 250 in Massachusetts and Central Connecticut; and 185 m.y. in New Hampshire and possibly Rhode Island.

It is evident that the most prominent event over the entire region occurred

TABLE 2

Potassium-Argon Age Measurements in the Plutonic-Metamorphic Province of the Southeastern United States

Sample Number	Sample Weight	A^{40*}	$\frac{A^{40*}}{\text{Total } A^{40}}$	K	$\frac{A^{40*}}{K^{40}}$	Apparent Age	Rock Unit	Location
	gm	ppm		%		m.y.		
Hoot Owl[a] M	20.7979	0.225		8.69	0.0213 ± 0.0011	334 ± 15	pegmatite	Spruce Pine, N. C.
McKinney[a] M	30.6736	0.214		8.24	0.0213 ± 0.0011	334 ± 15	pegmatite	Spruce Pine, N. C.
	25.9922	0.224			0.0223 ± 0.0011	348 ± 15		
Minpro[a] M	20.9981	0.238		8.97	0.0218 ± 0.0009	341 ± 15	pegmatite	Spruce Pine, N. C.
L-30 M + B	18.9193	0.208	0.93	7.77	0.0218 ± 0.0009	341 ± 13	Carolina gneiss	Adjacent to McKinney pegmatite
L-141 B[b]	11.2705	0.207	0.95	7.42	0.0229 ± 0.0009	357 ± 13	Carolina gneiss	Rte. 19 near Stocksville, N. C.
BK-149[c] B	5.9978	0.101	0.90	2.92	0.0284 ± 0.0011	433 ± 15	Metadiabase dike	Near Red Hill, N. C.
BK-61[c] B	15.6561	0.054	0.73	1.48	0.0301 ± 0.0015	457 ± 21	Metadiabase dike	Near Red Hill, N. C.
L-146 B	12.7250	0.221	0.94	7.92	0.0229 ± 0.0009	357 ± 13	Cranberry gneiss	Roan Mtn., Tenn.
L-22[b] B	14.0610	0.200	0.92	6.02	0.0273 ± 0.0011	418 ± 16	Beech granite	Rte. 19E near Hopson, Tenn.
L-333 B	16.0685	0.285	0.95	6.63	0.0354 ± 0.0014	527 ± 18	Cranberry gneiss	Near Rte. 19E bet. Hopson and Hampton, Tenn.
L-332[b] B	15.0496	0.346	0.95	6.39	0.0444 ± 0.0018	640 ± 22	Cranberry gneiss	Near Rte. 19E bet. Hopson and Hampton, Tenn.
L-331 B	19.5960	0.378	0.97	6.87	0.0451 ± 0.0018	648 ± 22	Cranberry gneiss	Rte. 19E near Hampton, Tenn.
L-328 B	12.8000	0.346	0.93	6.03	0.0472 ± 0.0019	674 ± 22	Cranberry gneiss	Rte. 19E near Hampton, Tenn.
L-149[b] B	16.0633	0.473	0.97	6.68	0.0581 ± 0.0023	800 ± 27	Cranberry gneiss	Pardee Point, Tenn.
L-330 B	7.6769	0.391	0.93	6.55	0.0490 ± 0.0020	695 ± 23	Cranberry gneiss	Near White Rock, N. C.
L-136 M	14.5851	0.239	0.93	9.04	0.0217 ± 0.0009	340 ± 13	pegmatite	Bryson City, N. C.
L-139 M	12.2940	0.228	0.89	8.41	0.0223 ± 0.0009	348 ± 13	pegmatite	Gay, N. C.
L-140[b] B	5.6739	0.262	0.70	7.39	0.0291 ± 0.0012	438 ± 14	Carolina gneiss	Gay, N. C.
	3.8716	0.260	0.77		0.0284 ± 0.0011			

TABLE 2 (Continued)

Sample Number	Sample Weight	A^{40*}	$\frac{A^{40*}}{\text{Total } A^{40}}$	K	$\frac{A^{40*}}{K^{40}}$	Apparent Age	Rock Unit	Location
	gm	ppm		%		m.y.		
L-132 M	16.8249	0.209	0.94	8.36	0.0205 ± 0.0008	323 ± 12	Talledega fm.	McCaysville, Ga.
L-132 B	16.8169	0.188	0.94	6.96[d]	0.0222 ± 0.0009	347 ± 13		
L-133[b] B	14.1783	0.215	0.64	6.23[d]	0.0284 ± 0.0011	434 ± 15	Ocoee fm.	Rte. 64 near Ducktown, Tenn.
L-25 B	14.7658	0.201	0.92	6.94	0.0238 ± 0.0009	370 ± 13	Lynchburg gneiss	Rte. 460 near Bedford, Va.
L-16 B	20.0160	0.200	0.93	7.19	0.0228 ± 0.0009	356 ± 13	granite	Mt. Airy, N. C.
L-118 B	19.8736	0.205	0.79	7.48	0.0225 ± 0.0009	352 ± 13	granite	Rte. 29-70 near High Point, N. C.
L-119 M	18.5990	0.208	0.93	8.53	0.0200 ± 0.0008	315 ± 11	Carolina gneiss	Rte. 150 near Lincolnton, N. C.
L-120 P	15.7585	0.151	0.89	6.33	0.0196 ± 0.0008	309 ± 11	Gaffney marble	Gaffney, S. C.
L-124 B	19.0201	0.062	0.76	3.63	0.0140 ± 0.0006	226 ± 9	granite	Cayce, S. C.
	16.9041	0.064	0.55		0.0144 ± 0.0006	233 ± 9		
L-125 B	21.2888	0.142	0.91	7.56	0.0154 ± 0.0006	247 ± 9	granite	Elberton, Ga.
L-127 B	18.9359	0.164	0.90	7.73	0.0174 ± 0.0007	277 ± 10	Lithonia gneiss	Rte. 78 near Snellville, Ga.
L-128 M	19.8313	0.192	0.87	8.52	0.0185 ± 0.0007	294 ± 10	granite	Stone Mtn., Ga.
L-129 B	13.3156	0.181	0.92	7.60[d]	0.0195 ± 0.0008	309 ± 11	Carolina gneiss	Buford dam site, Ga.
L-41[e] B	14.2653	0.235	0.92	7.53[d]	0.0257 ± 0.0010	396 ± 14	Baltimore gneiss	Radnor, Pa.
Westmore[f] M	20.2555	0.197	0.86	7.36	0.0220 ± 0.0009	344 ± 13	granite	Near Westmore, Vt

* Indicates radiogenic component
M Muscovite
B Biotite
P Phlogopite
a Carr and Kulp (1957)
b Also Rb-Sr age (Table 1)
c Sample supplied by R. Wilcox (Wilcox and Poldervaart, 1958)
d Analyzed by D. Thaemlitz and C. O. Ingamells, Rock Analysis Laboratory, Univ. of Minnesota; others by Ledoux and Company.
e Sample supplied by W. H. Tomlinson
f Sample supplied by J. Dennis

in the range from 320 to 380 m.y. ago. This range of dates is not due to analytical uncertainty. It seems clear that the orogeny in Nova Scotia slightly preceded that in the Central and Southern Appalachians, and that pulses in Northern New Hampshire and the Baltimore area were still later. A distinctly younger metamorphic event at about 250 m.y. ago is recorded in the Middletown, Connecticut pegmatites, in many Massachusetts granites and possibly, at a somewhat later time, in the southeast Piedmont. It appears, however, that this epoch was not as widespread as the major one which occurred around 350 m.y. ago.

The present precision and accuracy of these geochronometric techniques makes it possible to date approximately the major metamorphic events of the Appalachian Province. It is not possible as yet to answer the fascinating and important question of how long these metamorphic periods were and whether they represent one or many pulses of differing duration. Certainly if the metamorphism in the basement is directly related to the uplifts that produced detrital sediments in the Allegheny Geosyncline to the west (Woodward, 1957), some activity must have been more or less continuous from the early Silurian through the Pennsylvanian. The age results suggest only two events of major proportions in post-Cambrian time centering at about 350 and 250 m.y. Many publications have correlated these dates with the Taconian and Acadian disturbances. A completely different interpretation is possible, however, which, although contradictory to the generally accepted Holmes' time sacle, appears to digest more of the rapidly accumulating age data. The more important facts are:

1) Hercynian rocks in the Black Forest Region have been dated by Faul (1957) by the Rb-Sr and K-A methods at about 335 m.y. These rocks are definitely pre-Westphalian (middle Carboniferous) but are usually assigned to the Dinantian (lowest Carboniferous) which on the Holmes' time scale is placed at 265 m.y.

2) The M.I.T. work in Nova Scotia has determined ages of 325 to 385 m.y. for rocks that intrude lower Devonian strata.

3) Work in this laboratory to be published elsewhere (Cobb and Kulp, in preparation) has shown that the Upper Cambrian Swedish kolm is at least 500 m.y. old.

If these facts are accepted as reliable it is clear that the Taconic orogeny cannot be placed at 350 m.y. In this regard it is interesting that Woodward (1958) suggests that "much of the Taconic deformation either should be backdated to the Trenton or that it took place in unison with the Caledonian of the Silurian. Relatively little active deformation in the northeastern Appalachian basin seems left for necessary assignment to the closing days of the Ordovician." It is also possible that there is no necessary correlation between metamorphic pulses in the basement and depositional or structural patterns in the geosyncline to the west.

It is suggested therefore that the most satisfactory solution appears to be a considerable lengthening of the Paleozoic Era putting the Acadian orogeny in the 380 m.y. range, i.e., Middle-Upper Devonian, and the 330 to 360 activity in the Southern Appalachians in the Mississippian during which time there

was uplift in the east of Tennessee and Alabama. The 250 m.y. event in the southeast might be correlated with the Appalachian orogeny at the end of the Paleozoic. The Taconian orogeny of Ordovician time is then earlier than these and in general obscured by the overriding events in the span of 320 to 380 m.y. ago. The top of the Cambrian would therefore be placed at about 500 m.y. ago.

CONCLUSIONS

1. The major Paleozoic metamorphic event now displayed in the plutonic-metamorphic complex of the Southeast Piedmont and Blue Ridge occurred about 350 m.y. ago, slightly later than the major Paleozoic orogeny in the Northern Appalachians which appears to have begun about 360 m.y. ago.

2. The primary basement in the Southeast is at least 900 m.y. old. In all probability its true age of formation is close to 1100 m.y., having formed at the same time as the Grenville Province of the eastern Canadian Shield.

3. In western North Carolina it is possible to pass through a transition zone 10 miles wide across the structural trend from rocks that have had their micas completely recrystallized by the 350 m.y. event to rocks farther northwest that have not been altered. This emphasizes the danger of using isolated age measurements in metamorphic terrains to indicate the actual time of a metamorphic or igneous event. If the isolated sample happened to come from a transition zone like that in westernmost North Carolina, the apparent age is only an index of the extent of recrystallization.

4. It cannot yet be decided whether the 350 m.y. metamorphism in the Southern Appalachians is to be correlated with the Acadian or younger Mississippian activity. The youngest events in the Southeast (~250 m.y.) may reflect the Appalachian orogeny at the end of the Paleozoic.

5. It seems certain that a distinct regional metamorphism occurred in the Southern Appalachians between the time of formation of the original basement (~1100 m.y.) and the 350 m.y. event. The possibility that this occurred in Lower Paleozoic time cannot be excluded.

Further isotopic age studies in the Southern and Central Appalachians are in progress at this laboratory.

APPENDIX

Description of rock specimens

Purpose: The following descriptions of thin sections of dated samples are intended to aid in evaluating the ages reported above. Therefore the descriptions take note primarily of those features which reflect the geologic history of the rock. These observations are of particular importance in view of the possibility that a rock may have undergone more than one cycle of metamorphism, or metamorphism subsequent to crystallization from a silicate melt.

Hoot Owl, McKinney, Minpro, *Fresh book muscovite, Spruce Pine pegmatite district, N. C.*: (See Carr and Kulp, 1957)

L-30, *Carolina gneiss adjacent to McKinney Mine, Spruce Pine pegmatite district, N. C.*: Olive green biotite and muscovite both occur as medium-grained flakes intergrown with each other. Chlorite is only present in occasional small areas, along with fine grains of iron ore, both forming at the expense of biotite.

Garnet, epidote, clinozoisite, iron ore, allanite, apatite and sphene are of local occurrence in association with the micas. Epidote is usually found rimming allanite. Quartz is strained; plagioclase occurs with and without twinning, being readily recognized by limited formation of sericite.

L-141, *Carolina-type gneiss, Rte. 19-23 near Stocksville, N. C.*: Medium- to fine-grained plagioclase crystals occur in a thoroughly granulated and finely crushed groundmass consisting mainly of interlocking grains of strained quartz. Fine- to medium-grained brown biotite is streaked through the rock, showing some alteration to chlorite, epidote and clinozoisite. Scattered grains of subhedral to anhedral sphene are present in the crushed groundmass. Some green biotite has formed at the expense of brown biotite.

"BK-149, *Coarse-grained metadolerite*: The average grain size of this subophitic rock is 2.0 mm. Dominant minerals include slightly clouded, lath-shaped plagioclase, in general twinned and with indistinct boundaries, and equidimensional augite, also slightly clouded and with irregular outlines. Degree of clouding in both augite and plagioclase varies in different crystals and within one individual. In plain light, 'channel-ways' apparently separate adjacent areas of clouding in a crystal. Subhedral to anhedral, pale-pink, anisotropic garnet, crowded with inclusions, is associated with, and may rim augite. Subhedral, isotropic garnet is in contact with plagioclase. Skeletal magnetite is scattered throughout the rock and in most cases is rimmed by garnet. Reddish-brown biotite is associated with magnetite or distributed throughout the rock. The few orthopyroxenes present follow the habit of augite. Tiny granules of greenish-brown hornblende rim some augite." (Wilcox and Poldervaart, 1958)

"BK-61, *Fine-grained metadolerite*: This rock contains relict plagioclase and pyroxene, recrystallized plagioclase and pyroxene, and biotite, garnet, and magnetite. Hornblende is present only in trace amounts. Relict plagioclase is slightly clouded and forms laths up to 1 mm long with ragged edges. Recrystallized plagioclase builds irregular, equidimensional crystals about 0.1 mm in diameter. A few plagioclase porphyroblasts up to 2 mm in diameter are slightly clouded and contain abundant inclusions. All plagioclase contains minute inclusions of biotite, pyroxene, and hornblende. Relict augite forms irregular, moderately clouded crystals. Clouding is concentrated along cleavage cracks. Recrystallized pyroxene granules may be associated with relict augite, or they may build granules and clusters of granules which are scattered throughout the rock. Light- to dark-brown biotite averages about 0.05 mm in diameter. Biotite flakes are mounted on magnetite, garnet, and recrystallized granoblastic pyroxene, and the mineral is therefore clearly of metamorphic origin. Garnet is pale pink, averages 0.1 mm in diameter, and is in general anhedral, although subhedral crystals are found in contact with plagioclase. It commonly surrounds magnetite and relict augite. Most garnets are choked with inclusions and appear anisotropic. Inclusions are principally minute pyroxene, magnetite, and biotite, along with abundant submicroscopic particles. Skeletal magnetite is scattered throughout the rock and in many places is surrounded by garnet. Scattered sparsely throughout the rock are tiny hornblende granules associated with recrystallized plagioclase, pyroxene, biotite,

and magnetite. The association garnet-hornblende is not common in this rock." (Wilcox and Poldervaart, 1958)

L-146, *Cranberry gneiss adjacent to metadiabase dikes, on Roan Mtn., 1.3 mi. north of Tenn.-N. C. line on Rte. 143*: Biotite is light- to dark-brown, fine-grained. The larger flakes are crenulated, locally shredded and disaggregated at the margins, giving rise to fine-grained biotite of similar color. Allanite is occasionally present in biotite flakes, producing pleochroic haloes. Rare grains of epidote are associated with biotite. Garnet is slightly pink, medium- to fine-grained, and sieved by quartz. Plagioclase is unaltered while quartz has undulatory extinction.

L-22, *Freshly blasted specimen of Beech granite from quarry on Rte. 19E near Hopson, Tenn.*: Microcline is coarsely perthitic; medium-grained plagioclase laths are bent and broken; and large areas of quartz have pronounced undulatory extinction. Granulation of mineral grains is common. Plagioclase shows some sericitization. Biotite is deep olive green, fine-grained; present in trace amounts due to nearly complete alteration to chlorite (showing limonite staining), clinozoisite, epidote and iron ore; and found with the above minerals exclusively in granulated portions of the rock. Poorly formed, clear garnet is present in the crushed areas.

L-333, *Cranberry gneiss, Tiger Valley Road, 4.3 mi. south of junction with Rte. 19E. (Junction is 5.5 mi. southeast of Hampton, Tenn.)*. Microcline and plagioclase are medium-grained. Plagioclase shows fairly extensive alteration to sericite. Microcline is perthitic. Biotite is brown to olive green, bent, crenulated and often shredded in appearance. Iron ore forms large elongate patches in association with fine-grained epidote, clinozoisite and olive green biotite. Discrete grains of iron ore occur throughout and marginal to biotite flakes. The larger brown biotite flakes have been altered marginally to an olive green biotite, which in many cases has broken down to a large number of fine-grained flakes peripheral to the main areas of mica. Chlorite containing rutile needles has locally formed at the expense of biotite. Some granulation of quartz and feldspar is seen, the former having undulatory extinction.

L-332, *Cranberry gneiss, 2.0 mi. south of Rte. 19E on Tiger Valley Road*: Plagioclase and microcline are present with the former showing more extensive sericitization. Biotite is fine-grained, brown; individual flakes are bent and altered to chlorite and iron ore. Some muscovite occurs peripheral to and replacing biotite. Quartz forms large patches showing undulatory extinction.

L-331, *Cranberry gneiss, 5.8 mi. southeast of Hampton, Tenn. on Rte. 19E*: Plagioclase is medium grained, is often untwinned and generally sericitized. Biotite is brown in color; locally shows extensive alteration to chlorite; margins and cracks are emphasized by abundant grains of iron ore. Muscovite rims many biotite flakes and appears to have formed at the expense of the latter. Quartz consistently shows undulatory extinction.

L-328, *Cranberry gneiss, 3.8 mi. south of Hampton, Tenn. on Rte. 19E.*: Biotite is medium- to fine-grained; brown to dark brown, to olive green, with occasional grains altered to chlorite. Medium- to fine-grained microcline is the dominant feldspar, plagioclase is subordinate and moderately altered to sericite. Quartz is strained, occurs as large irregular patches.

L-149, *Cranberry gneiss collected near contact with overlying unmetamorphosed sedimentary cover, Pardee Point, Tenn.*: Considerable feldspar is untwinned; all feldspar is moderately sericitized, alteration being most intense along fractures. Quartz is strained, occurring in elongate areas parallel to the general trend of biotite flakes. Reddish brown biotite makes up 10-20 per cent of the rock; occasional flakes are completely altered to chlorite and iron ore.

L-330, *Cranberry gneiss, 2.2 mi. southwest of White Rock, N. C.*: All feldspar is completely sericitized and beyond identification. Quartz grains show strong undulatory extinction, are fractured and veined by carbonate material which forms discrete grains elsewhere. Biotite is reddish brown, very thoroughly crenulated on a fine scale, contains fine grains of iron ore, and in places has completely broken down to a fine-grained aggregate of light tan biotite and muscovite. Grains of iron ore are common, some being rimmed by sphene. Occasional chlorite is seen marginal to and within biotite flakes. Subhedral apatite is fairly common.

L-136, *Pegmatitic muscovite, Deep Creek No. 1 Mine, Bryson City, N. C.*

L-139, *Pegmatitic book muscovite, small pegmatite, Gay, N. C.*

L-140, *Carolina-type gneiss, Rte. 23 at Gay, N. C.*: Medium- to fine-grained garnet porphyroblasts occur in a fine-grained groundmass of strained quartz, brown biotite and twinned plagioclase. Garnet appears to be replacing biotite which is associated with fine- to medium-grained grains of iron ore. No alteration of plagioclase feldspar to sericite, or biotite to chlorite is observed.

L-132, *Shear zone in Talledega formation, ~1.5 mi. southeast of McCaysville, Ga. on Rte. 245*: No thin section description available.

L-133, *Ocoee formation, 4.5 mi. west of Ducktown, Tenn., on Rte. 64*: Quartz forms small porphyroblasts with nearly uniform extinction. Most of the mica is fine-grained, consisting of approximately equal amounts of brown biotite and muscovite, which together make up 15 to 20 per cent of the rock. Iron ore occurs in streaks, elongated parallel to the mineral foliation, and as scattered grains. Occasional discrete grains of chlorite exist but do not appear to be an alteration of biotite. Fine-grained albite grains are present in the groundmass.

L-25, *Lynchburg gneiss, Rte. 460 between Lynchburg and Bedford, Va., 5 mi. east of junction 460-122*: Biotite is fine-grained with a preferred orientation, and in association with sphene, pink garnet, apatite and zoisite. Epidote and clinozoisite are rare, iron ore lacking. Apatite forms grains up to 1 mm long, is rimmed by sphene or zoisite, with the latter also present along cracks. Garnets are euhedral to anhedral, and vary from being clear of inclusions to highly sieved by minute foreign grains. Fine-grained biotite and zoisite are often intimately intergrown. Sphene is largely anhedral, forming coarse irregular patches. Paligoclase is rare and the sole feldspar. Quartz is abundant as interlocking fine grains forming irregularly shaped coarse areas in the groundmass. Quartz-rich bands alternate with bands richer in the other minerals mentioned. Occasional slight undulatory extinction is observed in the quartz.

L-16, *Granite, quarry, Mt. Airy, N. C.*: Biotite is olive green, occurs in

intimate association with fine-grained epidote, and often with sphene and apatite. Muscovite flakes vary from being well-formed to extremely shredded in appearance, the latter occurrences having an irregular surface appearance. Plagioclase and microcline are worm-like and medium grained, the former showing faint oscillatory zoning. Both feldspars are generally free of sericite. Undulatory extinction is faint in some quartz, absent in other large areas. Myrmekite is present. Epidote occurs rimming allanite, as does sphene in some instances. Chlorite has completely replaced occasional flakes of biotite.

L-118, *Freshly blasted so-called "carboniferous granite", Superior Stone Company quarry, 1/2 mi. south of Rte. 29-70, 6 mi. east of High Point, N. C.*: Biotite is deep olive green; fine-grained; and locally in close association with muscovite, sphene, iron ore, and epidote. Plagioclase shows only local light development of sericite while all quartz grains have uniform extinction.

L-119, *Clear, coarse book muscovite taken from alaskite body in deeply weathered Carolina gneiss, ~2 mi. south of Lincolnton, N. C. on Rte. 150.*

L-120, *Gaffney marble, quarry, Gaffney, S. C.*: The rock consists primarily of calcite. Mica is largely light tan phlogopite, plus some muscovite. Grains of sphene, rutile, iron ore, and tremolite are present.

L-124, *Granite, quarry, Cayce, S. C.*: Biotite has a shredded, disaggregated appearance, much of it having been completely altered to chlorite, iron ore, epidote and clinozoisite. Muscovite flakes occur marginal to the altered biotite. Occasional grains of allanite are observed in association with epidote. Sodic plagioclase is the dominant feldspar with microcline present in minor amounts. Quartz is strained and the groundmass records some granulation.

L-125, *Granite, Elberton City Quarry, Elberton, Ga.*: Microcline is the dominant feldspar with sodic plagioclase present in subordinant amounts; both feldspars are nearly free of sericite. Quartz extinction varies from uniform to undulatory. Biotite is fine-grained, marginally replaced by muscovite, the latter also occurring as separate flakes. Allanite rimmed by epidote is common in the rock and often occurs replacing the biotite. Iron ore, zircon and apatite are also present.

L-127, *Lithonia gneiss, quarry 4 mi. east of Snellville, Ga. on Rte. 78*: Biotite is fine-grained, deep olive green in color, often partly or completely rimmed by muscovite. Muscovite is very pale green in color, occurs both in association with and distinct from biotite flakes, and often has an irregular surface appearance. Sphene grains occur both in clusters and marginal to iron ore. Epidote and clinozoisite appear to be alteration products of biotite. Grains of allanite are rimmed by epidote and both are associated with biotite. Microcline and plagioclase are almost devoid of sericite; quartz usually shows uniform extinction.

L-128, *Granite, Stone Mtn., Ga.*: Fine-grained muscovite is common throughout the rock with only minor amounts of brown biotite present, the latter always occurring in association with muscovite. Occasional flakes of muscovite are bent, while some biotite is extensively altered to chlorite. Microcline and plagioclase are medium-grained and nearly void of sericite; quartz has nearly uniform extinction. Epidote and clinozoisite are present in association with the muscovite.

L-129, *Carolina-type gneiss, Buford dam site, near Buford, Ga.*: Biotite is brown to deep olive green, fine-grained and associated with equally coarse muscovite. Both minerals have a strong preferred orientation as do elongate areas of strained quartz. Quartz is the dominant mineral, plagioclase the sole feldspar. Garnet, apatite, epidote and iron ore occur as discrete grains scattered through rock.

L-41, *Baltimore gneiss, quarry, Radnor, Pa.*: Hypersthene and hornblende are fine- to medium-grained and clouded, while the latter is altered to fine-grained brown biotite at the margins and along cleavages. Fine-grained subhedral to anhedral garnets form nearly continuous rims around coarse-grained pyroxene, amphibole and biotite, and consistently occur at boundaries between plagioclase and the ferromagnesian minerals. Amphibole and pyroxene are present in approximately equal amounts with the former appearing to result from replacement of the latter. In thin section the rock has a streaked appearance due to lens-shaped areas of coarse plagioclase.

Westmore granite, near Westmore, Vt.: No thin section description available.

TABLE 3

K-A and Rb-Sr Dates on Micas in South Carolina and Georgia

Locality	Rock Unit	K-A	Rb-Sr	Laboratory
Cayce, S. C.	granite	230±10	—	Lamont (this paper)
Elberton, Ga.	granite	245±10	—	Lamont
Elberton, Ga.	granite	—	250±10	M.I.T. (Pinson et al., 1958)
Stone Mtn., Ga.	Stone Mtn. granite	295±10	—	Lamont
Stone Mtn., Ga.	Stone Mtn. granite	—	283±15	M.I.T.
Snellville, Ga.	Lithonia gneiss	275±10	—	Lamont
Mt. Arabia Quarry, Ga.	Lithonia gneiss	—	288±10	M.I.T.
Buford Dam Site, Ga.	Carolina gneiss	310±11	—	Lamont
Ben Hill, Ga.	Carolina gneiss	—	342±34	M.I.T.

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REVIEWS

Elements of Geology; by JAMES H. ZUMBERGE. P. viii, 382; figs. 215. New York, 1958 (John Wiley and Sons, Inc., $5.50).—The author set out to design a textbook for a one-semester course covering both Physical and Historical Geology. The resulting product is adequate and forms a reasonable framework for such a course. The book is easy to read, clear and straightforward in its explanations, and well illustrated by both photographs and diagrams. Use of technical terms has been kept to a minimum, and those used are for the most part adequately and clearly defined. However, the definition given for cleavage of minerals (p. 38-39) is poor and might lead the unsuspecting student into thinking that cleavage results in thin flexible layers, as in mica. In some instances the definition is not stated explicitly, but the reader is left to formulate his own from a list of requirements which must be filled by the definition.

The book is divided into 17 chapters; the first 11 are concerned with Physical Geology. In the introductory chapter the emphasis is laid on the concept of geologic time and on geology as a science in which field work is important. The importance of the concept of geologic time is brought out effectively several times elsewhere also.

The discussion progresses from the Earth's setting in the universe to the Earth's crust, and from the geologic forces at work on the crust to the specific agents and processes. The main topics discussed are earthquakes, rock weathering, windwork, riverwork, glacial sculpture, and shorelines. Into the chapter on the crust of the Earth the author has condensed the discussion of minerals, igneous rocks, sedimentary rocks, metamorphic rocks, and intrusions.

In the 6 chapters on Historical Geology the author proceeds from a general discussion of geologic maps, the meaning of sedimentary rocks, including the factors that control sedimentary rock properties, paleogeography, fossils, and the geologic time scale to the several eras and their periods. On pages 208-209 Zumberge attributes the Doctrine of Uniformitarianism to Lyell, failing to mention Hutton's roll in its formulation.

There is a very brief discussion of the history of each period and of the fossils found in the rocks of that age. At the end of the discussion of each era, the location and origin of economically important products formed during the era is discussed. It is in these discussions that the origin of coal and petroleum, as well as other products, is taken up.

Several typographic errors were noted, the most glaring being that the Sierran batholith is attributed to the Triassic on page 313 but to the Jurassic on page 297.

In the illustrations of the fossils the generic names are given for the several Pennsylvanian plants, but they are omitted in the illustrations of the invertebrates. It seems that the book might be more meaningful if these names were supplied.

Each chapter is opened with a brief introductory paragraph explaining the nature of the material to be covered. At the end of each chapter is a summary of its salient points; these summaries should be extremely useful to the student.

The author has prepared a sound textbook for the type of course for which it is designed, and although there are places where one might use a different plan of organization, might choose different topics, or might emphasize other points, the book is one that should be useful.

CHARLES W. WELBY

The Stratigraphy of Western Australia; by J. R. H. McWHAE, P. E. PLAYFORD, A. W. LINDNER, B. F. GLENISTER, B. E. BALME. P. 161; 19 figs., 1 map. London and New York, 1959 (Melbourne University Press; New York, Cambridge University Press, $8.50). *The Geology of South Australia*; edited by M. F. Glaessner and L. W. Parkin. P. 163, 26 figs., 11 pls., 5 maps, London and New York, 1959 (Melbourne University Press; New York, Cambridge University Press, $8.50).—A wealth of data has accumulated since the publication of David's impressive work, *The Geology of the Commonwealth of Australia* in 1950, and the authors and editors of these two new books have made valuable contributions to Australian geology by documenting and summarizing this data. Both volumes were originally published in the Journal of the young and vigorous Geological Society of Australia, which is to be congratulated on its achievement.

The Stratigraphy of Western Australia follows the outline of David's Volume 1 on historical geology with description of systems by areas. J. R. H. McWhae has carefully compiled the manuscript and illustrations with the assistance of the co-authors. The paleographic maps for each system and the stratigraphic columns are extremely informative. The interesting occurrences of fossiliferous Cambrian and Ordovician strata in north-east Western Australia, and the widely distributed upper Devonian, Permian, Cretaceous, and Tertiary strata are lucidly described and indicate the mass of data which has in part been gained by the recent search for oil. Correlation of the various formations has many problems and an immense field of research on the rich faunas awaits future investigators. The concluding section on tectonic history reviews the structural developments and igneous activity in Western Australia.

The Geology of South Australia, illustrated with excellent geological maps and cross sections, emphasizes the vast Pre-Cambrian Shield areas and offers great interest to the petrologist. The geology of the State is described in a series of geological provinces in which stratigraphic successions, tectonic development, and economic geology in each province are outlined in addition to the comprehensive study of Pre-Cambrian rocks. The four important Mesozoic and Tertiary basins are treated at length in relation to the tectonics. The extremely useful volume is a compilation of chapters prepared by authors with specialized knowledge of the various areas concerned.

These two books provide the most up-to-date and comprehensive summary of geological progress in Western Australia and South Australia. They are important reference books in Australian geology, building upon the secure foundation established by David's stock-taking and interpretation.

J. R. P. PHILLIPS

Prehistoric Animals; by JOSEPH AUGUSTA, illustrated by Zdenek Burian, P. 47; 69 pls. London, 1957 (Spring Books, $7.95).—This volume reproduces

a series of restorations of prehistoric animals that is notable for the careful representation of anatomical details, lifelike postures, and reasonable environmental surroundings of the extinct animals. Aside from a few introductory plates of early Paleozoic invertebrates and some restorations of marine reptiles, the volume is devoted to the land vertebrates. Each plate is accompanied by a full page explanation of the relationships, inferred habits, geologic occurrences and other interesting data concerning the animals and plants illustrated. The introductory chapter outlines the highlights of the geologic history of animal and plant life.

Some details of the restorations might be improved; the small size of the fish pursued by *Mesosaurus* (pl. 13) gives an errouneous impression of the size of that small reptile; *Mastodonsaurus* (pl. 15) appears too deep-bodied; *Brontosaurus* too low at the shoulder; the nostrils of many of the dinosaurs appear too small for creatures of such large size. Several color plates present refreshingly light scenes and depict the ancient animals with patterns that can never be substantiated from the fossils, but that are nevertheless plausible by analogy with living vertebrates.

The text is for the most part accurate and lucid, though in places the English translation leaves something to be desired. Although some recent discoveries and ideas have necessarily been omitted, it would serve reasonably well for an introductory text in the geologic history of vertebrates, and can certainly be recommended as a collateral reference in historical geology.

JOSEPH T. GREGORY

Pleistocene Man at San Diego; by George F. Carter. P. xvi, 400; 96 figs. Baltimore, 1957 (The Johns Hopkins Press, $8.00).—Archaeological research has pushed the time of man's arrival in the New World back as far as the Wisconsin glaciation. This book, by a Johns Hopkins geographer, would push it back still farther into the preceding, Sangamon interglacial. The author marshalls data in support of his conclusions from an impressively large number of specialties, including geomorphology, pedology, and climatology as well as archaeology. However, there is some reason to believe that his earliest remains, consisting of "hearths," "refuse," and "stone artifacts," are the product of natural rather than human agencies, as in the case of the Old World eoliths. Moreover, the interglacial dating of the terraces in which these remains have been found rests upon a number of unproven assumptions, e. g., that sea-level changes in the San Diego region are entirely eustatic. The author deserves credit for posing the problem of interglacial man in the New World but cannot be said to have resolved this problem.

IRVING ROUSE

Elementary Seismology; by CHARLES F. RICHTER. P. viii, 768; 205 figs. San Francisco, 1958 (W. H. Freeman and Co., $12.00).—This is the longest and most complete of the "all inclusive" modern treatments of seismology. Professor Richter has deliberately, and I think wisely, limited his treatment of certain subjects such as the description and extended theory of seismographs; however, virtually all aspects of the study of earthquakes are treated at least briefly.

The selection of topics and placing of emphasis should make this book of great interest to geologists. As could be expected, being the father of the earthquake magnitude scale which bears his name, Professor Richter presents an excellent discussion of magnitude and energy. The section on geography and geology of earthquakes is perhaps the greatest contribution in this book. This section makes up nearly a third of the book and will be of particular interest to anyone interested in regional tectonics.

There are seventeen appendixes making up over a quarter of the book. Most of these are of interest only to seismologists but many of us are finding them of much use. Included in these appendixes are lists of seismological stations, lists of large earthquakes, various tables and charts, summaries of formulas, etc.

There is a certain unevenness in the choice of material that may worry some readers, for example, the inclusion of a descriptive chapter on seismology in the USSR, an appendix on an African earthquake, and notes from the IUGG meetings in Toronto. I find this unevenness more than made up for by the author's engaging style and evident love for his branch of science.

The book is printed excellently, and is profusely and well illustrated. There are numerous tables and extensive lists of references. Professor Richter and his publisher are to be complimented.

JAMES T. WILSON

Advances in Petroleum Chemistry and Refining, Volume I; edited by KENNETH A. KOBE and JOHN J. MCKETTA, JR. P. vi, 641. New York, 1958 (Interscience Publishers, Inc., $13.50).—This book consists of 12 chapters, each of them a brief progress report on the present status of the subject concerned. One or more specific topics are chosen from each of the following areas: economics and future trends, unit operations and design, refining processes, petrochemicals, mechanical equipment. It is intended to present further volumes in this series at annual intervals, choosing specific topics in such a way that the entire industry will be covered over a period of a few years.

The authors of the various sections of this first volume are well qualified to write on their topics. The quality of the presentations ranges from good to excellent.

CHARLES A. WALKER

Publications Recently Received

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